ENERGY EXCHANGE IN THE BIOSPHERE

DAVID M. GATES

HARPER REPRINT SERIES IN PLANT PHYSIOLOGY—VOL. III

A HARPER INTERNATIONAL STUDENT REPRINT

A COMPLETE & UNABRIDGED REPRINT
(*U. S. list price $4.00*)

HERE is a biological monograph which spans the study gap between the biological world and the physical processes of the universe. This book is primarily a consideration of the interaction of biological organisms with their environment in a *quantitative* fashion, rather than in the traditionally qualitative style so often occurring today. Having found the traditional method inadequate, the author advances what, in his opinion, is the correct application of such factors as annual rainfall, sunlight, wind, air temperatures, etc., for describing an ecological environment. In the author's own words, the stated purpose of this monograph is "to provide a discussion of the energy regime of environments and of the interaction of plants and animals with the energy exchanges."

Pointing out energy exchange as the one predominant feature of the environments of living organisms, the author emphasizes its importance as an extremely critical condition for the existence and propagation of life itself, thus supporting his argument for the critical need for the proper integration of physics and mathematics into the study of environmental biology.

The energy factors the text deals with specifically are solar radiation; infrared; thermal radation from the atmosphere, the ground, and surrounding objects; the transfer of heat by convection, specifically in

(*continued on back flap*)

In India

P. Jarvis

Digitized by the Internet Archive
in 2022 with funding from
Kahle/Austin Foundation

https://archive.org/details/energyexchangein0000davi

Energy Exchange in the Biosphere

HARPER REPRINT SERIES
IN PLANT PHYSIOLOGY

ENERGY EXCHANGE IN THE BIOSPHERE

David M. Gates

Consultant to the Director
Physics of the Atmosphere
National Bureau of Standards
Boulder, Colorado

jointly published by
HARPER & ROW, New York, Evanston & London
and JOHN WEATHERHILL, INC., Tokyo

This Harper International Student Reprint is an unabridged, photo-offset reproduction of the American edition. It is for sale in Japan by John Weatherhill, Inc., 50 Ryudo-cho, Azabu, Minato-ku, Tokyo; in the United Kingdom and on the continent of Europe by Harper & Row, Ltd., 69 Great Russell Street, London W. C. 1; and in other designated countries by Harper & Row, Publishers Incorporated, 49 East 33rd Street, New York, N.Y. 10016 / Copyright © 1962 by Harper & Row, Publishers, Incorporated / All rights reserved; no part of this book may be used or reproduced in any manner whatsoever without written permission except in the case of brief quotations embodied in critical articles and reviews. For information write Harper & Row at the New York address given above / Library of Congress catalog card number 62-18882 / Printed in Japan / First reprint edition, October, 1965.

Preface

This book contains material which was given as a series of lectures to a seminar at the University of Minnesota during October, 1960. This seminar was jointly sponsored by the departments of Botany, Entomology, and Soils. The first lecture, Chapter 1, was presented as a Sigma Xi address on the St. Paul Campus and represented the keynote lecture for the seminar. The material was also presented as a course in the Biology Department of the University of Colorado during the autumn semester of 1961.

The lectures have been revised and a substantial amount of material has been added in order to give the subject matter the necessary depth and breadth for its printed form. As complete a bibliography as possible has been given in order that the reader will be encouraged to explore beyond the contents of this book. This is particularly true with regard to Chapter 3 on radiation instruments in which the references will show details of construction which could not be presented here.

The author particularly wishes to thank Professor Donald B. Lawrence, University of Minnesota, for initiating these lectures, for encouraging the publication of them, and for criticism of the manuscript. The author also wishes to express his appreciation to the following persons for their criticisms and suggestions: Professor George R. Blake and Dr. Donald G. Baker, University of Minnesota; Professor Frank Kreith and Professor John W. Marr, University of Colorado. The author assumes responsibility for the correctness of the material presented. He sincerely hopes that this work will assist with bringing into modern biology, especially ecology, more frequent use of the techniques and principles of the physical sciences, for it is indeed a basic fact that ecology, by definition, is truly interdisciplinary.

David M. Gates

Boulder, Colorado

Contents

Preface v

List of Figures vii

List of Tables viii

1. The Energy Environment in Which We Live 1

2. Solar and Thermal Radiation 47

3. Radiation Instruments 74

4. Heat Transfer by Convection 94

5. New Methods in Ecology 113

References 143

Index 149

LIST OF FIGURES

1. Energy Exchange of the Earth and Atmosphere for the Northern Hemisphere 5
2. Daily Totals of the Undepleted Solar Radiation for Different Geographical Latitudes 8
3. Extraterrestrial Solar Flux, the Solar Flux at the Earth's Surface, and the Flux from 6000°K Black Body 9
4. Idealized Temperature Profiles in the Ground and Air . . 11
5. Energy Exchange for an Animal in a Natural Environment 13
6. The Infrared Transmission of the Earth's Atmosphere to Infrared Wavelengths, the Estimated Atmospheric Thermal Emission Downward at the Surface, the Ground Emission Upward, and the Observed Atmospheric Thermal Emission 15
7. Map of the Annual Values of the Net Radiation at the Earth's Surface 22
8. Map of the Monthly Values of the Net Radiation at the Earth's Surface for the Month of June 24
9. The Annual Variations of the Heat Balance Components for Various Locations at the Earth's Surface 26
10. The Estimated Maximum Heat Load Received on the Upper and Lower Surfaces of a Horizontal Leaf in Various Environments 28
11. Variation in Solar Radiation and Environmental Temperature and the Resultant Radiant Heat Load on Man . . . 34
12. Radiosonde Air Temperatures Compared with Radiant Sky Temperatures Measured at the Zenith 36
13. Effect of Fog and Snow on Environmental Radiant Temperatures 38
14. Hourly Totals of the Net Radiation in a Stand of Pine Trees in Bavaria 39
15. The Solar Spectrum at the Mean Solar Distance from the Earth as Received Outside of the Earth's Atmosphere on a Surface Perpendicular to the Solar Rays 49
16. The Electromagnetic Spectrum and Means of Detection . . 61
17. Spectral Radiancy of Black Body Sources at Various Temperatures 62
18. Stern Pyranometer . . . . . *facing* 78
19. Bellani Pyranometer 80
20. Franssila-Suomi Net Radiometer or Balance Meter *facing* 78

21. Hofman Heat-Compensated Net Radiometer or Balance Meter . . . *facing* 79
22. Stoll-Hardy Radiometer . . . 87
23. Schulze Radiometer . . . *facing* 79
24. Wind Velocity and Air Temperature Profiles near a Plane Surface . . . 95
25. The Wind Velocity Profile above a Very Rough Surface . . 109
26. The Total Absorption in the 2.7, 4.3, and 15 Micron CO_2 Bands . . . 119
27. The Total Absorption in the 1.38 Micron H_2O Band . . . 121
28. The Total Absorption in the 1.87 Micron H_2O Band . . . 122
29. The Total Absorption in the 2.7 Micron H_2O Band . . . 123
30. The Total Absorption in the 3.2 Micron H_2O Band . . . 124
31. The Total Absorption in the 6.3 Micron H_2O Band . . . 125
32. The Spectral Characteristics, Absorptivity, Reflectivity, and Transmissivity of Ficus Leaves . . . 138
33. Solar Intensity at the Earth's Surface at 14,000 Feet Altitude . . . 139
34. Solar Intensity at the Earth's Surface at Sea Level . . . 140

LIST OF TABLES

I. Cloud Cover Parameters for Solar Insolation as Influenced by Various Cloud Types . . . 56
II. Ratio of Insolation with Overcast Sky to Insolation with Cloudless Sky for Various Cloud Types . . . 57
III. Values of R_c . . . 69
IV. Vapor Pressure Values . . . 69
V. Radiation Regimes for Certain Hypothetical Environments . . . 70
VI. Radiation Instruments . . . 75
VII. Typical Values of Viscosity for Water and Air . . . 97
VIII. Nusselt Numbers for Free or Natural Convection . . . 100
IX. Convection Coefficients for Free or Natural Convection in Laminar Flow . . . 101
X. Reynolds and Nusselt Numbers and Convection Coefficients for Forced Convection across a Flat Plate . . . 105
XI. Rate of Heat Transfer for Forced Convection across a Flat Plate . . . 106
XII. Rate of Heat Transfer for Forced Convection across a Cylinder . . . 107

Chapter 1

The Energy Environment in Which We Live

INTRODUCTION

Environments have been of great interest to scientists for many years and have been described and measured in a variety of ways. The study of environments has been the concern of meteorologists, climatologists, geographers, geologists, botanists, zoologists, agriculturalists, foresters, and others. Although the author is a physicist, who has been working on problems of atmospheric physics, he has long had a serious interest in the subject of biology and nearly entered this branch of science rather than physics. Since the atmosphere is resting on the biosphere, so to speak, and since organisms not only spend their whole lives interacting with one another but also interacting with their immediate physical environment, it is not surprising that this physicist should take a renewed interest in the biological problems. Looking at the various environments ranging from tropical rain forests, deserts, lake shores, grasslands, boreal forests, alpine meadows to arctic tundras, one is struck by one predominant feature besides the characteristics of vegetation and animal life, and that feature is the energy exchange. The flow of energy into and out of these environments must in many ways be an extremely critical condition for the existence and propagation of life itself. Energy and the exchange of energy from one object to another in the laboratory, industry, the atmosphere, the stars, and now in outer space have been of great interest to the physicist, the astronomer, and the engineer for many years. The subject of heat transfer has been thoroughly developed and is now one of the most esteemed subjects in most engineering schools. The application of this branch of physics to biology is long overdue. It is now essential and critical that physics and mathematics be properly integrated into the study of environmental biology.

To begin with, one point should be made emphatically clear. Biology is by far a more difficult subject than physics. It is much more complex, and it has more variables. It is more difficult to do a controlled experiment and to understand the basic laws of nature which apply in biology than it is in physics. The complexity of problems in biology means that the subject should be attacked with all the skills and tools available. One cannot separate the biological world from the physical processes of the universe. To study biology without simultaneously reaching an understanding of physical mechanisms is to superficially isolate the biological problem from reality. Biologists should be trained as thoroughly as possible and in many disciplines, including mathematics, physics, and chemistry, as well as the traditional subjects of biology. This may require more years of study than is now expected; however, there is little choice. Modern science is capable of advancing biology far faster than it is now moving. Those students who will educate themselves thoroughly in the basic sciences will reap a rich harvest in the field of biological research.

The world in which we live is indeed a complex one. Conditions on the planet earth happened to have been just right for life to have evolved. It may be a fairly delicate balance, a rather thin thread upon which we dangle. It is tremendously important to consider the interaction of biological organisms with their environment in a quantitative fashion, rather than in the traditionally descriptive qualitative style so often occurring today. I wish to introduce in the following pages a quantitative approach, which, in my opinion, is the way to study environments and the relationships of organisms to them.

It has been traditional for the biologist to describe an ecological environment in terms of its plant and animal association and such meteorological factors as annual rainfall, sunlight, air temperature, wind, and moisture, as well as soil characteristics. However, it should be remembered that some of these factors are perfectly good parameters for describing the properties of the atmosphere, but if they are to be used for describing an ecological environment, then they must be correctly applied. Very rarely in the ecological literature are such data properly utilized. Usually the data are amassed and the statement often made, "little correlation with plant growth was noted" or "it was difficult to find a correlation." Actually, in fact, it would have been most surprising to find a high degree of correlation in many instances. The parameters air temperature, soil temperature, dew point, solar radiation, moisture, and wind speed are indeed important; they are important not in themselves, but for the purpose of describing the energy regime and energy exchange of the environment. One must learn how to use them and when and where they are required; their use is rarely simple. It is the purpose of these lectures to provide a discussion of the energy regime of environments and of the interaction of plants and animals with the energy exchanges.

THE FLOW OF ENERGY

Sunlight, thermal radiation, evaporation, convection, and conduction all interact with an organism on the surface and either deliver a net energy to the organism or abstract energy from it. The following equation expresses the various components of energy as they stream to and from the surface of the earth in the boundary layer in which most terrestrial organisms spend their lives.

$$S + R + LE + G + C + s = 0 \tag{1.1}$$

where S = solar radiation flux; R = thermal radiation flux from the ground, atmosphere, and surrounding objects; L = latent heat of evaporation; E = rate of evaporation; G = sensible heat flux by conduction from the ground through the soil or rock to the surface; C = sensible heat flux by convection in the air; and s = a storage term for storing energy or giving up energy over short periods of time. The term "flux," when applied to radiant energy, refers to the total amount of radiant energy of all wavelengths crossing a unit area of surface per unit time, in units of cal per cm^2 per min. When applied to the movement of water or mass, the term "flux" refers to the amount crossing a unit area per unit time in units of grams per cm^2 per min.

For living organisms, the components are essentially the same, except the reflective, transmissive, and emissive properties of the organism must be taken into account for both visible and infrared wavelengths. The equation is now written:

$$S - (r + t)S + \varepsilon R + LE + G + C + s + P + M = 0 \quad (1.2)$$

where r = reflectivity, t = transmissivity, ε = emissivity, P = photosynthesis, and M = basal metabolism. The organism may exchange energy like a gray body, therefore, radiating a fraction, ϵ, of the energy radiated by a black body. The term S again represents the total solar energy incident upon the organism, and the term $(r + t)S$ represents the energy unavailable to the organism through reflection and transmission losses. The term $S - (r + t)S$ represents the solar energy absorbed by the organism. The latent heat term remains as before, the ground conduction term may remain if the organism is in contact with the ground, the convective term is almost always present, and then there will be basal metabolism and possibly also photosynthesis going on within the organism itself.

Figure 1 shows mean values for the various components of energy as they pertain to the atmosphere of the northern hemisphere of the earth. If one considers 100 units of solar radiation passing through a horizontal surface at the top of the atmosphere, these 100 units will represent 0.485 cal per cm^2 per min. This is the annual mean value in the northern hemisphere of the solar radiation energy incident upon a horizontal surface located outside the atmosphere. Of these

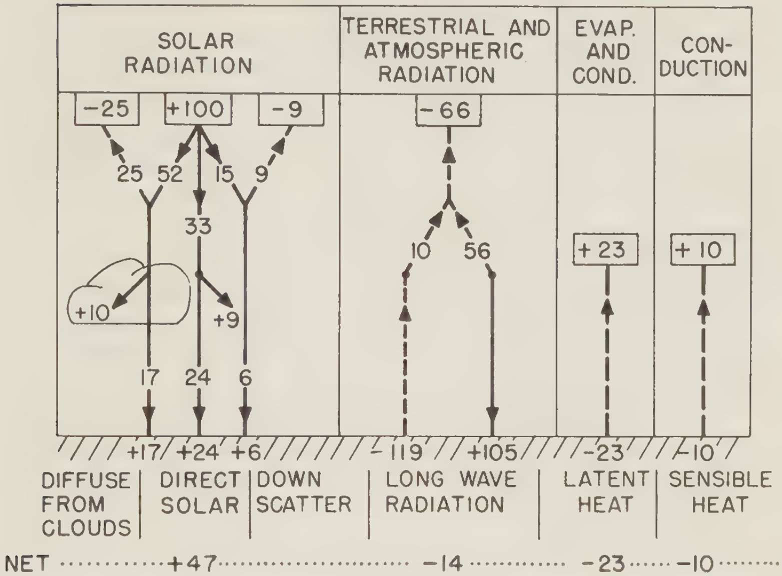


FIGURE 1.—Energy exchange of the earth and atmosphere for the northern hemisphere (100 units = 0.485 cal per cm^2 per min). Based on the solar constant value of 1.94 cal per cm^3 per min.

100 units, one can consider that, on the average, 52 units will be dealt with by the clouds in the atmosphere. The clouds will reflect 25 units back to space, absorb 10 units themselves, and permit 17 units to stream to the ground to be available at the surface. On the average, 24 units will be transmitted directly through clear skies to the surface, 6 units are scattered by the sky to the surface. The total energy available at the ground will average 47 units or 0.228 cal per cm^2 per min on a horizontal surface. Actually, of course, these exact numbers are not encountered in practice, but represent the average of both summer and winter conditions for the northern hemisphere to give the energy budget of the atmosphere and surface.

The earth and the atmosphere radiate heat in all directions. This is one of the phenomena which will be discussed in detail later. The warm surface of the earth radiates heat very effectively, very much like a black body. Of 119 units radiated outward from the surface, 10 units escape to space and 109 units are absorbed in the atmosphere. The atmospheric gases reradiate this energy and send a stream of 105 units to the surface and 56 units to the cosmic cold of outer space. The atmosphere acts like a gray body; it has a temperature gradient with height, and it radiates heat. This process is always occurring, day or night, summer or winter, sunny or overcast. This heat radiated by the earth's atmosphere, both upward and downward, is long-wave infrared radiation with energy distributed throughout the infrared spectrum but reaching a maximum at a wavelength of about 10 microns. The next energy exchange term is the very important evaporation-condensation process which takes place at the surface. On the average, the earth's surface in the northern hemisphere is feeding 23 units of energy into the atmosphere by means of the evaporation process. This moisture comes from the soil, from the lakes and rivers, and from the plants growing on the surface. This energy is not lost from the earth because the moisture that is vaporized at the surface is recondensed in the form of clouds in the atmosphere, thereby giving back 23 units of latent heat to the atmosphere. Therefore, as shown in Figure 1, there is a minus 23 units at the surface and plus 23 units higher up. The last term to be considered in the energy exchange of the earth-atmosphere system is the sensible heat component. Sensible heat is the heat transferred by convection because of the wind flowing across the surface of the earth. In Figure 1 it is labeled conduction, but it is actually conduction-convection together. The winds across the surface carry away 10 units and deposit those 10 units in the atmosphere. This simply means that on the average the surface of the earth is warmer than the air, and the air is gaining heat by mass flow over the surface.

And so it is that Figure 1 represents the energy flow between the earth's surface and the atmosphere where the organisms are spending their entire lives. It is important to emphasize again and again that every living organism, you and I, the other animals, the plants on the face of the earth, spend their entire lives in a thermal radiation, infrared heat exchange environment which may be modified by solar radiation, convection, and latent heat exchange. In some circumstances, such as the arctic winter, there is no solar radiation, and the organisms are existing entirely in a thermal heat radiation environment. You and I in a building are exchanging radiation with the walls, and our comfort and survival depends upon this radiation field. We are not dependent upon the incandescent lights or visible light. We are quite comfortable in the dark, with the walls exchanging heat with us according to the surface temperature of the walls and ourselves. Here is a good example of a strong radiational environment. This environment is not modified by wind, so your comfort and your adjustment to this environment do not depend upon the sensible heat term. Nor is there much evaporation taking place, although our exhaled air and perspiration will put some moisture into the room while we are sitting there. Food must be periodically consumed in order to maintain the internal energy and body heat against the losses by radiation and evaporation.

The solar radiation term, the S of Equations 1.1 and 1.2, will now be described in some detail. The amount of energy per unit time from the sun incident upon a surface placed outside the earth's atmosphere perpendicular to the sun's rays at the mean distance of the earth from the sun is called the "solar constant." This has been a very difficult quantity to measure because of the disturbing influence of the atmosphere and the broad wavelength distribution of the energy. The best value is now thought to be 2.00 cal per cm^2 per min (see Johnson, 1954). Previously the value was thought to be 1.94 cal per cm^2 per min, and Figures 1 and 2 are based on this value. It is interesting to see how the amount of solar energy incident on a horizontal surface varies over the globe as a function of latitude and season. Figure 2 gives the daily totals of undepleted solar energy as received by a horizontal surface outside of the earth's atmosphere. The tremendous differences in the amount of solar energy flux at various positions on the earth are immediately evident. At Libreville, on

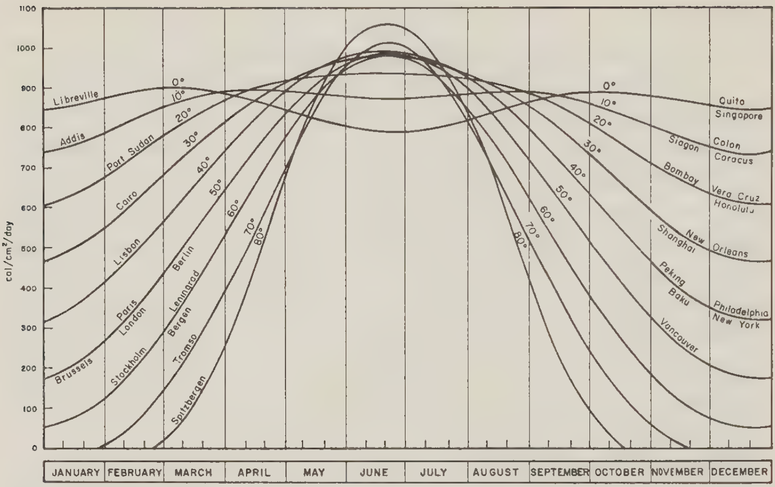


FIGURE 2.—Daily totals of the undepleted solar radiation received on a horizontal surface for different geographical latitudes as a function of the time of year. Based on the solar constant value of 1.94 cal per cm² per min.

the equator, the changing position of the earth in its orbit around the sun has very little effect throughout the year. The solar radiation environment of equatorial regions is very constant and hence is one of the reasons for the relative constancy of the energy regime of equatorial regions as is evidenced by the extremely small seasonal variations in soil temperatures, among other things.

Farther north in temperate latitudes, such as New Orleans, Washington, Paris, London, and Berlin, it is evident from Figure 2 that there is a considerable seasonal variation in the total daily solar energy incident upon a horizontal surface. Still farther north the variation in the daily totals becomes more extreme from summer to winter until finally north of the Arctic Circle the direct winter sunlight becomes zero in midwinter and the skylight illumination becomes small. At Spitsbergen there is no direct sunlight for nearly three months, only twilight, and the environment is composed of thermal radiation, modified at times by turbulent winds.

The solar radiation has to penetrate and diffuse through

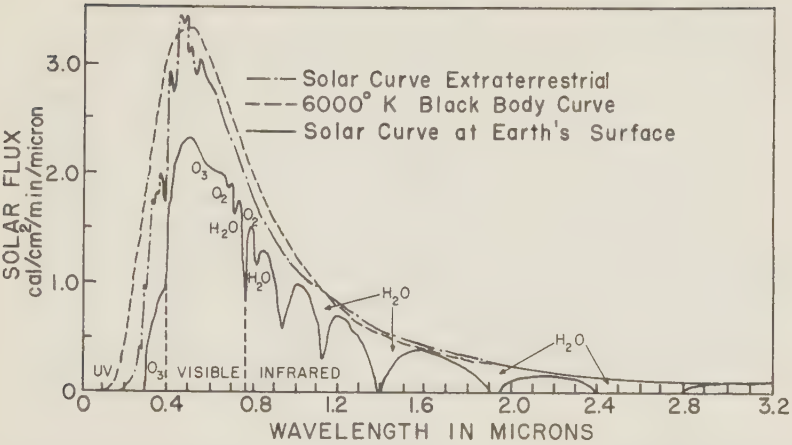


FIGURE 3.—Extraterrestrial solar flux, the solar flux at the earth's surface, and the flux from 6000°K black body as a function of the wavelength. The absorption bands resulting from atmospheric O_2, O_3, and H_2O are shown.

the earth's atmosphere, and when it gets to the surface it has been strongly depleted by means of absorption at discrete wavelength positions and by a continuum of scattering at all wavelengths. Figure 3 shows the spectral distribution of the solar energy when it reaches the surface as well as outside of the earth's atmosphere before it has been attenuated by scattering and absorption. For comparison purposes the emission from a 6000°K* black body at the mean distance of the sun is also shown. The visible portion of the spectrum is only a fraction of the total energy radiated by the sun. The solar energy curve peaks in the green of the visible at 0.48 microns. On the ultraviolet side it is depleted strongly through absorption by ozone and oxygen, and on the infrared side through absorption by water vapor, carbon dioxide, and ozone. The powerful actinic rays in the ultraviolet would "burn up" most organic material if it were not for the protecting layer of ozone up in the stratosphere and the oxygen molecules distributed to much greater heights in the atmosphere. The absorption of light by atoms or molecules is due to the fact that the molecule or atom resonates at definite frequencies corresponding to energy transitions between quantized energy states. Each molecular or atomic species will have a unique set of absorption lines or bands by means of which the absorbing constituent can be identified. In this way we know precisely the nature of the absorbing gases in the atmosphere. You will notice many absorption bands in the infrared attributed to water vapor and a strong depletion of the solar energy curve by these bands. Some biological textbooks state that the solar energy curve ends at 3 microns. This is quite incorrect, because, as discussed later, some sunlight does come through beyond this wavelength, although it is indeed the tail of the curve and represents a relatively small contribution to the total energy.

---

* Kevin scale, formerly called Absolute temperature, degree units of same size as the Celsius or Centigrade scale, but with zero 273 degrees lower.

Let us look for a moment at what is happening during the day and night at the ground, upon which all these organisms are dwelling, with regard to the temperature regime. Figure 4 is a schematic illustration of the temperature profiles in the air above and in the ground below the surface at midnight, after sunrise, at noon, and before sunset. The horizontal scale represents temperature with increasing values to the right, and the vertical scale is a measure of the height above and below the surface. By dawn the ground surface has cooled off to some low value, and this temperature is shown at the far left of the figure. The air above and the ground below are both warmer than the surface and so the temperature profiles slope to the right. At some depth well below the surface, about 30 centimeters, there is no diurnal temperature change, and all profiles must meet at this point. The annual wave of heat will penetrate to a depth of about 600 centimeters. The rate of penetration for any heat wave in the soil is about 2.2 cm per hr, the exact rate depending on the thermal conductivity of the soil under consideration and the temperature gradient. The direction of the energy fluxes due to ground conduction, *G*, turbulent convection, *C*, and latent heat of evaporation or condensation, *L*, are shown by arrows and by the symbols $>$ or $<$, representing "greater than" or "less than," respec-

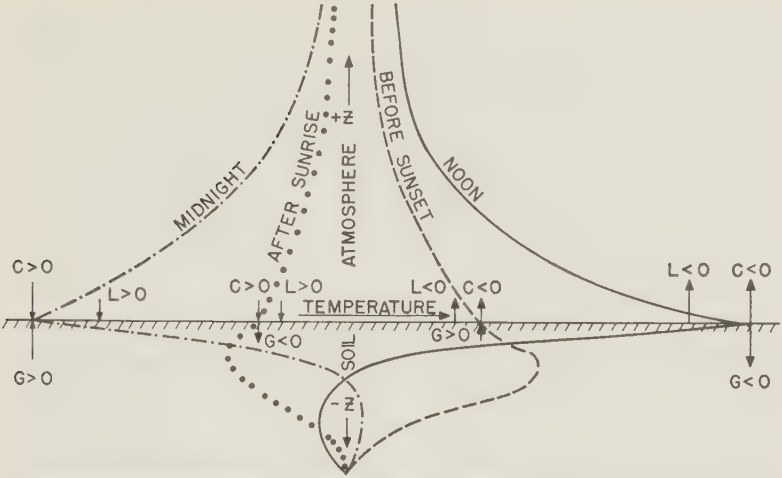


FIGURE 4.—Idealized temperature profiles in the ground and air for various times of the day, and the transport of heat by convection, *C*, by conduction, *G*, and by latent heat of evaporation or condensation, *L*.

tively. All quantities will be positive if heat is flowing toward the surface and negative if flowing away from the surface. By dawn, with the surface cooler than elsewhere, the energy flow due to convection, conduction, and latent heat must be toward the surface, from the air downward, and within the ground upward. Condensation may take place at the surface. Soon after sunrise the surface has been warmed somewhat by the sun. However, there is a lag beneath the surface so that the temperature drops as one penetrates a few inches into the soil, and then the temperature must reverse and end up at the nodal point at sufficient depth where all profiles meet. Above the surface the air is still warmer than at the surface, and the usual nocturnal temperature inversion still prevails. Heat must be conducted away from the surface into the ground toward the lower temperature strata and from the air toward the surface from above. Evaporation at the surface has probably begun at this time of day. By noon or early afternoon, one can picture the surface having been warmed to its maximum extent, and this point is shown at the far right in the diagram. The ground and the air are both distinctly cooler than the surface, and heat must be conducted and convected away from the surface. Evaporation must now be quite strong providing there is moisture available. The ground warms the air near the surface. This air rises because of its buoyancy, turbulence results, moisture is carried into the air, dust rises aloft, visibility goes down, winds come up, and the disturbed conditions of midday are in full swing. In the late afternoon, before sunset, the temperature of the surface has dropped but is still warmer than the air. However, there is a distinct temperature lag in the ground, and the temperature at a few inches depth remains warmer than the surface. In the ground, heat must be conducted toward the surface and toward deeper levels, while in the air, heat must be convected away from the surface. Some evaporation is probably still occurring. From these four examples it is easy to project the temperature profiles for other times of the diurnal cycle. It is in this boundary layer, near the surface, subjected to the diurnal changes of the heat flow, where large numbers of organisms spend their entire lives. The deeper roots of plants are in a rather undisturbed energy regime. Those roots which penetrate deep into the soil, say more than two feet, exist in an extremely constant thermal environment. Those roots nearer the surface have the harshness of the temperature changes at the surface considerably softened. The vegetation above the surface may undergo temperature excursions which are much greater than the surrounding air. For this reason it is important not only to understand qualitatively the pulsating dynamics of this environment at the surface, but to appreciate thoroughly its quantitative significance as well. One cannot consider what happens on the bare exposed ground as completely analogous with what happens above and below a ground covered with vegetation. However, the qualitative picture shown here is illustrative of the energy flow at the surface and the extent to which it may vary diurnally. For a vegetated surface the temperature profiles in the air may have quite different slopes than those shown here, and in fact one knows that the profile will change quite abruptly on passing from vegetation into free air above. In the daytime the air in the canopy of trees may be quite warm compared to the free air above them. At night the air in the canopy may be cooler than the air above the trees, due to the radiative cooling of the leaves themselves.

Figure 5 illustrates the heat load with which an animal living on the surface of the earth must contend. The animal is out in the open and is being irradiated with sunlight—direct sunlight, scattered sunlight from the sky, coming in from all directions. He gets reflected light from the ground, the leaves of the plants, the trees, and everything around him, including any clouds in the sky. He receives thermally radiated infrared radiation from the atmosphere, clouds, ground, and surrounding vegetation. He, in turn, is a hot radiating object, reradiating to space, to the sky, to the ground, and thereby losing heat by radiation in all directions. There is convective loss—turbulent transfer due to the air flowing across his body. This loss will occur whether or not there is wind. Without wind it is called natural convection; with wind it is called forced convection. The convective loss

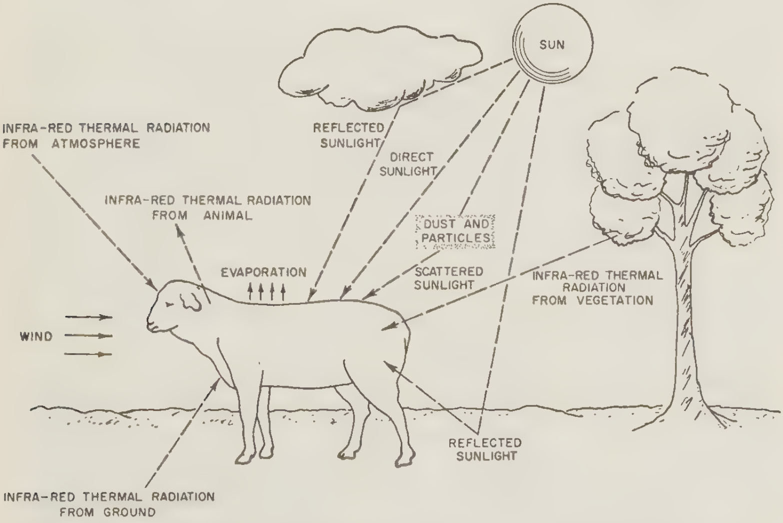

SUN
INFRA-RED THERMAL RADIATION FROM ATMOSPHERE
REFLECTED SUNLIGHT
DIRECT SUNLIGHT
INFRA-RED THERMAL RADIATION FROM ANIMAL
DUST AND PARTICLES
EVAPORATION
SCATTERED SUNLIGHT
INFRA-RED THERMAL RADIATION FROM VEGETATION
WIND
REFLECTED SUNLIGHT
INFRA-RED THERMAL RADIATION FROM GROUND

Figure 5.—Energy exchange for an animal in a natural environment.

can be substantial depending upon the organism and the nature of its body surface. The animal may sweat and hence there will be an evaporative loss of moisture.

The earth's surface radiates infrared radiation to the atmosphere and through the atmosphere to outer space, and the atmosphere itself emits infrared radiation downward to the ground and outward to space. While the solar radiation flux strikes the earth only during daylight, the long-wave, infrared streams of energy are always present, day or night, winter or summer. All organisms emit radiation and are, in turn, subjected to a constant flux of thermal radiation. These fluxes of infrared radiation have been generally ignored by the biologist in environmental studies although they actually represent a very important component of the energy exchange. It is the purpose now to describe the features of this long-wave, infrared heat exchange as it pertains to organisms living on the surface of the earth.

The atmosphere is composed of 78% nitrogen, 21% oxygen, and 0.93% argon by volume for dry air. Minor constituents which make up the remaining fraction include, among others, carbon dioxide, ozone, nitrous oxide, and methane. These minor constituents are widely variable in the biosphere. The carbon dioxide concentration may vary considerably within vegetated regions, but over the globe as a whole it averages about 0.03% by volume. Ozone at the surface is generally considered to be 4 parts per 10 million parts of air, but increases by tenfold or more in city smog conditions. Ozone also increases with altitude where its peak concentration of about 6 parts per million occurs in the stratosphere at 25 km above the surface. Nitrous oxide, which is a product of bacterial action in the soil, has a concentration near the surface of about 5 parts per 10 million parts and is quite variable. Water vapor, which is the most variable constituent over the surface of the earth, has concentrations from 0.1% to 1% by volume for most air. The concentrations given above were all for dry air. These molecular constituents possess numerous absorption bands—regions where they absorb incident energy—in the infrared. The regions of transparency between the absorption bands are termed "atmospheric windows." Some of the absorption bands due to water vapor and carbon dioxide (see Figure 3) cut off at a wavelength of 3.2 microns. However, beyond that wavelength, toward longer wavelengths, there are many more very important absorption bands due to water vapor, carbon dioxide, and ozone. These are shown in Figure 6. The transmission of the atmosphere to solar radiation of wavelengths greater than 4.0 microns is shown at the top of the figure. It should be noted that the solar spectrum as received at the earth's surface extends as far as 23.0 microns before finally being terminated. Solar radiation penetrates through the atmosphere again at very long

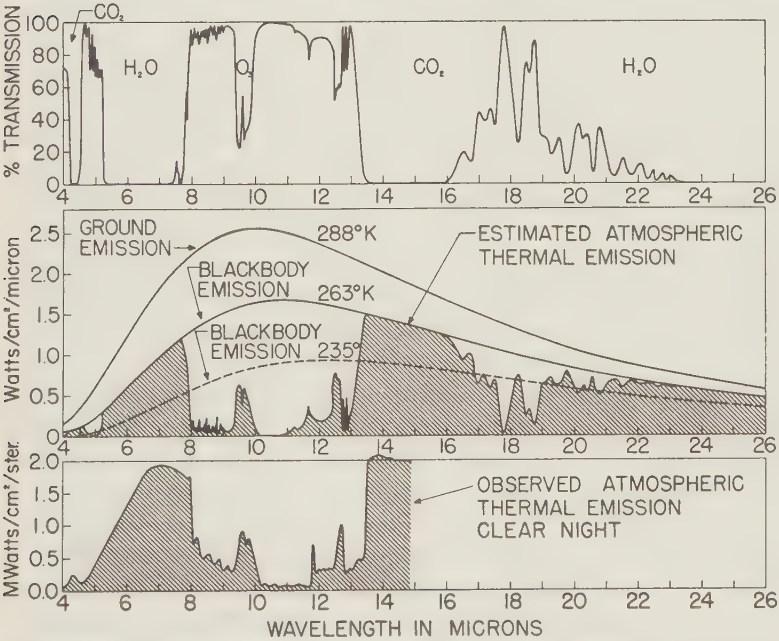


FIGURE 6.—The infrared transmission of the earth's atmosphere to infrared wavelengths, the estimated atmospheric thermal emission downward at the surface, the ground emission upward, and the observed atmospheric thermal emission. The emission from black body sources is at the mean tropospheric temperature of 263°K and the mean stratospheric temperature of 235°K.

wavelengths in the microwave or radio region at about 1000 microns or 1.0 millimeter. The absorption bands due to water vapor, located at 6.3 microns and from 18 microns and beyond, those associated with carbon dioxide at 4.3 and 15.0 microns, and the 9.6 micron band of ozone are extremely important from the standpoint of radiation transfer in the atmosphere. The reason for their importance lies in the fact that they are located in the wavelength region where radiators at the temperature of the earth or atmosphere emit most of their energy, namely 4.0 to 50.0 microns.

If the earth's atmosphere was composed of only oxygen, nitrogen, and the inert gases, which make up over 99% of the atmosphere, and did not contain the trace constituents water vapor and carbon dioxide, this planet would be a most uninhabitable place indeed. The gases nitrogen, oxygen, argon, etc., have no significant absorption bands in the infrared. They transmit sunlight readily but in addition transmit freely the long-wave infrared radiation from the surface of the planet to outer space. The earth would be hot on the daytime side and intensely cold on the nighttime side, having a climate not too dissimilar from that existing on the moon. It is doubtful from the purely radiation standpoint whether any form of life could exist under such conditions, and certainly true that higher forms of life could not have evolved. The truly minor constituents, water vapor and carbon dioxide, produce the critical control on the radiation exchange between the earth, the sun, and outer space which mollifies the climate at the surface to give the biosphere tolerable temperature limits.

The physics of thermal radiators is extremely well known and can be found in any elementary physics textbook. However, it is important for the present discussion to review the basic characteristics of radiators. A black body is defined as an object which will absorb completely all frequencies of radiation incident upon it. A good black body is constructed by cutting a small hole in the wall of a cavity or enclosure. The hole or opening indeed appears black, and all energy incident upon it will be trapped inside. This black body cavity will also emit radiation of all wavelengths back out through the hole in the wall. The radiation emerging represents the maximum energy which can be emitted at the temperature of the cavity. All other radiators or absorbers are compared to this black body. The surface of the earth, particularly soil, can be considered as a good approximation to a black body radiator. The first fundamental law of radiation states that the total energy emitted by a black body is proportional to the fourth power of the absolute temperature (degrees Kelvin). The second law of radiation states that the wavelength maximum for the emitted energy distribution multiplied by the absolute temperature is a constant. For a black body at the temperature of the earth, e.g., 288°K, the emitted energy will peak at nearly 10 microns wavelength. This curve is shown in the center of Figure 6 and is labeled "ground emission." Notice how the energy is distributed from a wavelength less than 4.0 microns to wavelengths off the diagram well beyond 26 microns. The total energy emitted will be the area under the curve. If the earth were at a higher temperature, it would emit energy according to a curve above the one shown and shifted toward shorter wavelengths. Notice that in Figure 3 a black body radiator at a temperature of 6000°K, approximating the solar emission, had its energy peak in the green of the visible at a wavelength of 0.48 microns. If the atmosphere were to absorb completely all wavelengths, then it too could be considered to be a black body. However, as has been pointed out, the atmosphere only absorbs energy in certain discrete absorption bands. If the atmosphere were a black body, it could be considered as radiating downward toward the surface as if it were at a temperature of about 263°K. This is shown in Figure 6 as the curve "black body emission 263°K." A third law of radiation states that a good absorber is a good emitter at the same frequency or wavelength. Since the atmosphere absorbs in discrete bands it will emit energy in these same discrete bands and to the degree to which they absorb. By applying the curve at the top of Figure 6 to the curve marked "black body emission 263°K," one gets the crosshatched areas marked "estimated atmospheric thermal emission." The only exception to this procedure is for the ozone band located at 9.6 microns. This is because most of the ozone is located in the stratosphere at a height of about 25 kilometers above the surface and must be considered as radiating at a temperature of 235°K. Applying the shape and strength of the 9.6 micron ozone band at the top of the figure to the curve marked "black body emission 235°K," one gets the shaded peak at 9.6 microns in the "estimated atmospheric thermal emission" curve. It is now interesting to compare the estimated curve with an actually observed night sky emission shown at the bottom of the figure. It is seen that all of the emission features are confirmed and in about the correct relative strength. There is a slight difference in the two vertical scales used since the report of the observations (Sloan, Shaw, and Williams, 1955) did not give sufficient information to permit a conversion.

Hence, one sees from Figure 6 that the atmosphere above does not flood the earth's surface with a continuous wavelength stream of energy, but sends down strong bands of energy located primarily between 5 and 8 microns and from 13 microns toward longer wavelengths, with smaller streams in between. It is of further importance to note that these bands of energy depend upon the atmosphere being composed of some water vapor and some carbon dioxide. The amount of carbon dioxide is 0.03% by volume and is well mixed over the globe although there are local inhomogeneities. Therefore the flux of energy in the region of 13 to 16 microns, which is due to CO_2, is reasonably constant over the globe and probably will not significantly reflect the small variations in CO_2 which occur in the biosphere. But the other fluxes, which are due to water vapor, vary tremendously with the water vapor content of the atmosphere over the globe, which is far from constant, being about 0.1% by volume for relatively dry air and 1% by volume for saturated warm air. Everyone has experienced the feeling of a clear sky on a dry, cold night and how much colder it feels than the air temperature would indicate. This is because the body is radiating heat and at the same time receiving relatively little from the atmosphere in return. If the moisture content of the atmosphere were to increase suddenly by a substantial amount, or if a layer, even a thin layer, of clouds were to move in over head, then a much stronger stream of thermal radiation would flood the surface and one would feel distinctly more comfortable. This very phenomena was witnessed in dramatic fashion on the evening of this lecture (24 October 1960, St. Paul, Minnesota). The author described what could happen if clouds moved in and how very cold it felt outdoors before the lecture because of the clear sky overhead. Unknown to the lecturer, during the evening a veil of clouds did develop over St. Paul, and when one stepped out of doors after the lecture the environment felt distinctly warmer. It is probably true that the air temperature had changed relatively little.

CARBON DIOXIDE THEORY OF CLIMATIC CHANGE

The carbon dioxide theory of climatic change has received considerable attention in recent years, notably through the publications by Callendar (1938, 1958) and by Plass (1956a, b,c, 1959). This theory is based on the fact that the fundamental absorption band of the CO_2 molecule is located in the infrared at 14 microns where the earth's atmosphere radiates heat efficiently. The carbon dioxide in the atmosphere effectively shields the earth's surface from the cosmic cold of outer space. An increase in the carbon dioxide content of the atmosphere would produce increased effectiveness of the shielding effect and by this means make the surface warmer. On the basis that the carbon dioxide content of the earth's atmosphere has increased by 13% during the past one hundred years, Plass calculated that the average temperature of the earth near the surface should have increased by about 1°F or 0.56°C. Since this is almost exactly the average increase recorded all over the world for the last century, it has been regarded as confirmation of the carbon dioxide theory of climatic change. Bray (1959) has restudied the question of the change in atmospheric carbon dioxide concentration during the last century and comes to the conclusion that the concentration was 294 ppm during the period 1857 through 1881 and 319 ppm during the years 1932–1956. Reports of analyses on carbon dioxide samples from many stations in Scandinavia are made periodically by Egnèr and Eriksson in the journal *Tellus* and may be found there. The amounts vary with the station and with the time of the year, the extremes running from 245 to 380 ppm.

Kaplan (1960) and Kondratiev and Nilisk (1960), among others, have had doubts concerning the magnitude of the effect of carbon dioxide on climatic change and have made extensive new calculations. According to Kaplan, halving of the carbon dioxide content of the atmosphere would reduce the mean temperature of the earth's surface by between 1.3°C and 1.9°C, and because of this it would seem that carbon dioxide variations could not play a major role in the ice-age cycle unless the carbon dioxide changes were by an order of magnitude. Kaplan further concludes that a change of 10% in the carbon dioxide concentration should result in a change of not more than 0.25°C in the surface temperature. Kondratiev also comes to the conclusion that the effect of carbon dioxide on climatic change is relatively insensitive. Kondratiev and Nilisk show in particular that the effect of water vapor absorption overlapping with the carbon dioxide absorption in the wavelength region 12 to 18 microns acts in such a way as to almost completely suppress any possible influence by carbon dioxide on climatic change through radiative processes.

There is no question but what the carbon dioxide theory of climatic change is an appealing hypothesis and that it should not be set aside as settled once and for all. Perhaps new evidence, either from the standpoint of the carbon dioxide content of the atmosphere, or with regard to the details of the radiation process, will someday make the hypothesis more valid. However, in the meantime, the carbon dioxide hypothesis should not be considered likely to be the primary reason for major climatic changes. The effects of cloud cover on climate will be far more dramatic than the possible effects by carbon dioxide. Also the effect of the mean water vapor content of the atmosphere, which is highly variable, may be of considerable influence. Calculations are required to show the magnitude of this effect for water vapor.

ENERGY REGIMES OF THE WORLD

The net radiation at the surface of the earth is the difference between all downward streams of radiation minus all upward streams of radiation. The downward flow of radiation includes direct solar, diffuse sunlight scattered by the clouds and sky, and thermal infrared radiation from the atmosphere and clouds. The upward flow of radiation includes reflected sunlight from the surface and thermal infrared radiation radiated by the surface of the ground. The net radiation at the surface, which is often referred to incorrectly as the radiation balance, is of great importance for consideration of the evaporation rates and the generation of thermal winds, and for the basic understanding of the energy budget at the surface. Budyko (1956) describes the world-wide distribution of net radiation as given by maps in the Radiation Atlas which he had published earlier (Budyko, 1955). The map giving annual totals has been redrawn and is reproduced in Figure 7. The energy units are kg-Cal per cm^2 per year. It is interesting to note some comparisons over the globe. The largest values of net radiation are found over the oceans where cloud cover is more prevalent, especially at night, thereby causing the surface to lose less heat to the cosmic cold of outer space, and at the same time the clouds are effective radiators themselves sending a strong flux of black body radiation toward the ocean surface emitted from their bottom sides. Although the increased cloud amounts over the ocean do somewhat reduce the amount of sunlight reaching the surface, this reduction is less than the increased blanketing effect producing more long-wave thermal infrared radiation as described above. The increased moisture content of oceanic air also contributes to the downward stream of infrared radiation. The greatest value of net radiation is 140 kg-Cal per cm^2 per year over the Indian Ocean where a very warm, moist at-

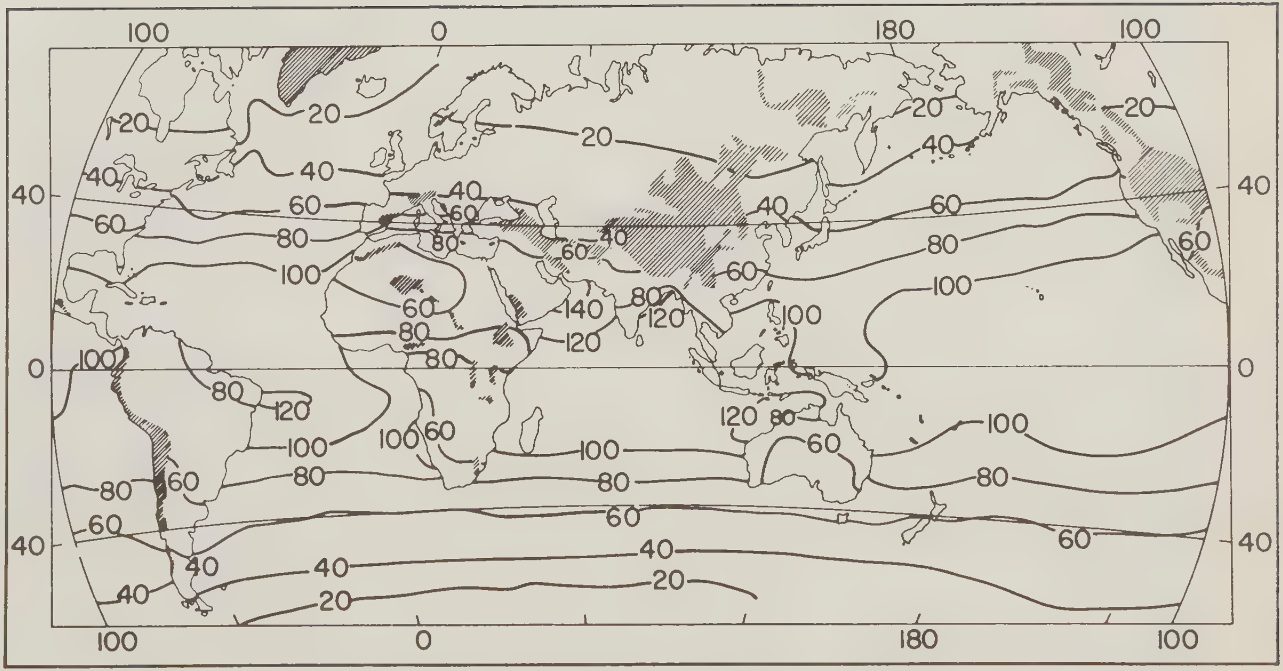


FIGURE 7.—Map of the annual values of the net radiation at the earth's surface in kg-Cal per cm² per year. This represents the net solar radiation at the surface plus or minus the net thermal radiation from the atmosphere and the ground surface. (Data from Budyko, 1955, 1956.)

mosphere prevails throughout much of the year. There are abrupt discontinuities between the oceans and the continents generally amounting to 20 to 40 kg-Cal per cm^2 per year. The continents may have more sunshine because of clearer skies, but they also have much stronger nocturnal cooling at the surface because of reduced cloudiness at night; in fact, the thermal radiation loss from the surface through the clear atmosphere to space will occur constantly day and night. Therefore, over the Sahara Desert where daytime temperatures may be extremely high due to the intense sunlight at the surface, the nighttime temperatures may be surprisingly low due to the nocturnal cooling by radiation loss through the very dry, transparent air. It will be noted from Figure 7 that the annual net radiation over the Sahara Desert is 60 kg-Cal per cm^2 per year. The value of the albedo of desert soils will also materially affect the net radiation values. Across northern United States through Minnesota the value is estimated to be about 40 kg-Cal per cm^2 per year. As one goes north the annual total amount of sunlight becomes reduced and the net radiation at the surface falls accordingly. Even arctic regions with continuous summer illumination have a small accumulation of positive net radiation throughout the year. The worldwide values for the month of June are shown in Figure 8. Here it will be noticed there are negative values in antarctic regions due to the lack of sunlight during the south polar winter. The net radiation values shown on these maps are averages, and large deviations from these can and do exist locally, due to orographic features affecting cloud distribution and humidity of the air, and to various soils and vegetations.

The author experienced the consequences of an unusually large net radiation at Locarno in Switzerland where the vegetation is unexpectedly tropical and characteristic of vegetation found seven hundred miles to the south in Sicily. The reason for this unusual condition is orographic since Locarno is located on the southern slopes of the Alps. This results in clear skies in the daytime, thereby permitting a great deal of sunshine to reach the surface, and the nights are frequently

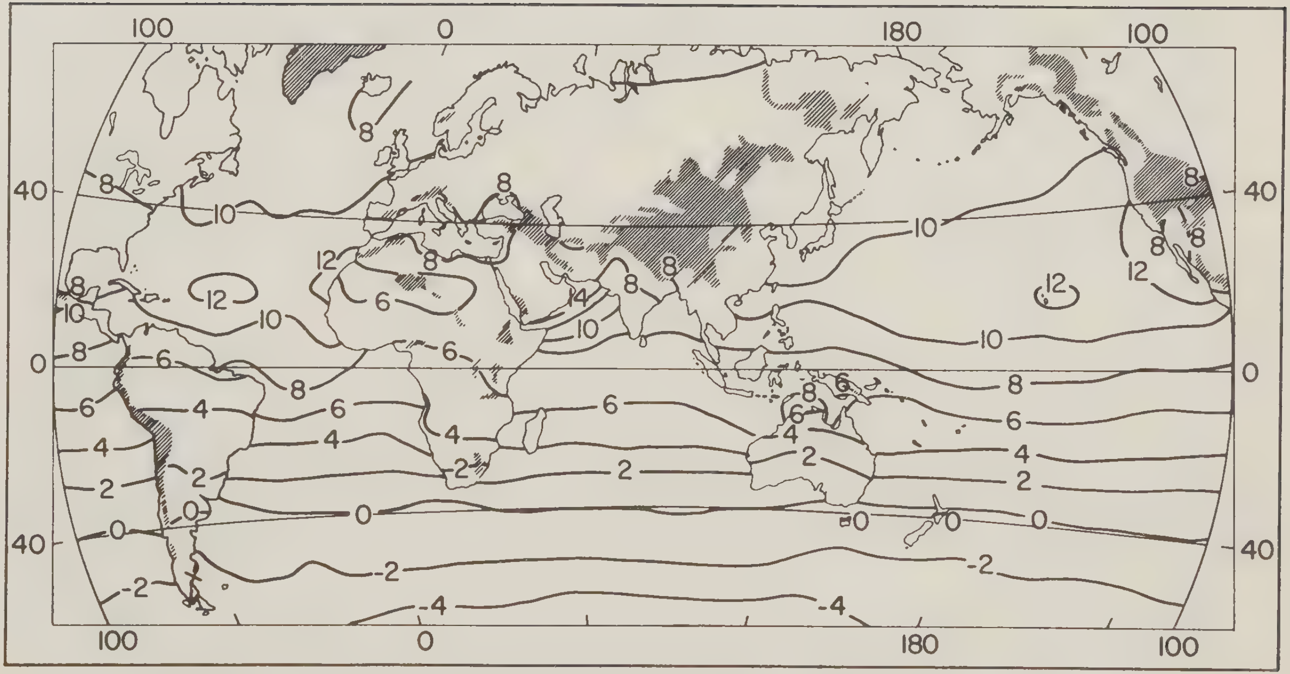


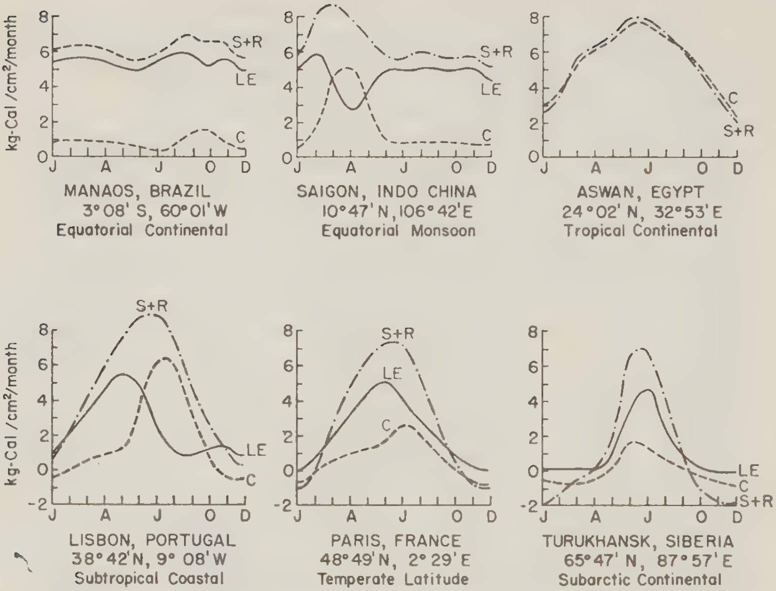
FIGURE 8.—Map of the monthly values of the net radiation at the earth's surface for the month of June in kg-Cal per cm² per month. This represents the net solar radiation at the surface plus or minus the net thermal radiation from the atmosphere and the ground surface. (Data from Budyko, 1955, 1956.)

cloudy with heavy rainfall. Locarno has the greatest amount of sunshine and the greatest annual rainfall in Switzerland. The vegetation type and growth are clearly a response to this heavy energy budget and moisture regime.

The net radiation at the surface, which results in a gain or a loss of energy, must either evaporate moisture, condense moisture, or generate air density gradients which will result in thermal winds. In the energy balance equation the net radiation represents the sum of the first two terms $S + R$. For conditions at the surface or just above the ground, the conduction term G can be neglected. Except for the storage of heat, the net radiation must be carried away by evaporation LE and convection C. In general the rate of evaporation will keep pace with the heat load where soil moisture is adequate. In this case the transfer of heat by convection will be small. When there is not sufficient soil moisture to maintain adequate evaporation proportional to the heat load, the energy must go into thermal winds and the convective transfer term becomes large. A comparison of climates throughout the world for various seasons of the year will illustrate this idea. In Figure 9 are shown the principal components of the energy regime for six locations on the earth. These examples represent certain extremes to be encountered as well as certain moderate conditions. In this diagram, contrary to the convention employed earlier, positive values of LE and C represent energy carried away from the surface.

At Manaos, Brazil, there is almost constant soil moistening by precipitation, and most of the net radiation is used in evaporation. The turbulent component is very small. The net radiation regime is nearly constant throughout the year. This is consistent with the constancy of the solar insolation in equatorial regions and the isothermal conditions of tropical soils as noted by Weber (1959).

For Saigon, Indochina, there is also a reasonable constancy of the net radiation with the exception of a sudden rise in late winter when the cloud cover present throughout the rest of the year suddenly diminishes and the insolation increases as a result. During this drier period evaporation diminishes and



MANAOS, BRAZIL
3° 08' S, 60° 01' W
Equatorial Continental

SAIGON, INDO CHINA
10° 47' N, 106° 42' E
Equatorial Monsoon

ASWAN, EGYPT
24° 02' N, 32° 53' E
Tropical Continental

LISBON, PORTUGAL
38° 42' N, 9° 08' W
Subtropical Coastal

PARIS, FRANCE
48° 49' N, 2° 29' E
Temperate Latitude

TURUKHANSK, SIBERIA
65° 47' N, 87° 57' E
Subarctic Continental

FIGURE 9.—The annual variations of the heat balance components in kg-Cal per cm² per month for various locations at the earth's surface. *S* + *R* is the net radiation at the surface, *C* is the heat transferred to or from the surface by convection or turbulence, and *LE* is the heat transferred by evaporation or condensation. (Data from Budyko, 1955, 1956.)

more of the net radiation must be carried away by convection. This period of the year in Saigon is characterized by more gusty winds. Throughout the remainder of the year the turbulence is very small indeed.

As one goes north from equatorial into tropical regions the net radiation begins to exhibit a strong seasonal variation. A typical dry, continental, tropical locality is Aswan, Egypt. Here the rainfall is so small as to have a negligible effect on the energy regime. The annual variation of the net radiation is considerable, and the turbulent exchange follows this curve closely. It should be noted that the maximum net radiation is not as large at Aswan as it is at Saigon, due to the higher albedo of the desert sands and the stronger outgoing infrared radiation from the warmer ground through the transparent dry air. The desert conditions are generally dry, dusty, and windy.

The Mediterranean climate is characterized by wet winters and dry summers, and the energy relationships for Lisbon, Portugal, exhibit this feature. The evaporation keeps pace with the net radiation increase in the spring, but then as the summer dry period sets in, the evaporation diminishes and the energy must be dissipated by convection, which reaches a maximum in July. It will be noted that the convection term is negative; this means that warm turbulent air, probably from over the sea, is delivering heat to the surface. This is the reason for the evaporation term being larger than the net radiation term during December. The heat delivered by the warm winds that blow must be carried away through evaporation.

For continental regions in temperate latitudes, there is a more even distribution of the energy flux among the components. Paris, France, represents a good example of this situation. Here once again winds that are warmer than the snow-covered ground deliver energy to the surface during November to February, and the evaporation amount is well above the net radiation level which is negative during this period. Throughout the remainder of the year the rainfall, and hence the evaporation, is not sufficient to discharge all of the net radiation accruing at the surface, and so there is always some turbulence with the winds stronger and more frequent during midsummer.

At Turukhansk, Siberia, the net radiation is negative—a loss of radiation from the surface—during the season mid-September to April, and the winds which are slightly warmer than the snow-covered tundra nearly compensate for the negative radiation regime. The evaporation is therefore zero during this same winter period. The summer peak in the net radiation shows up as an increase in convection until the snows melt, and then a peak is reached in evaporation while the winds have diminished in their role to carry away heat.

LOCALIZED RADIATION REGIMES

Having surveyed the world-wide energy relationships, it is now important to evaluate some specific localized conditions. Figure 10 is designed to exhibit and compare the solar and thermal radiation load of some specific situations. The lower curve in the diagram shows the average flux of sunlight, both direct and scattered, received on a horizontal surface in the latitude zone 40° to 50° for various times of the year. The upper curve shows the average flux of extraterrestrial solar

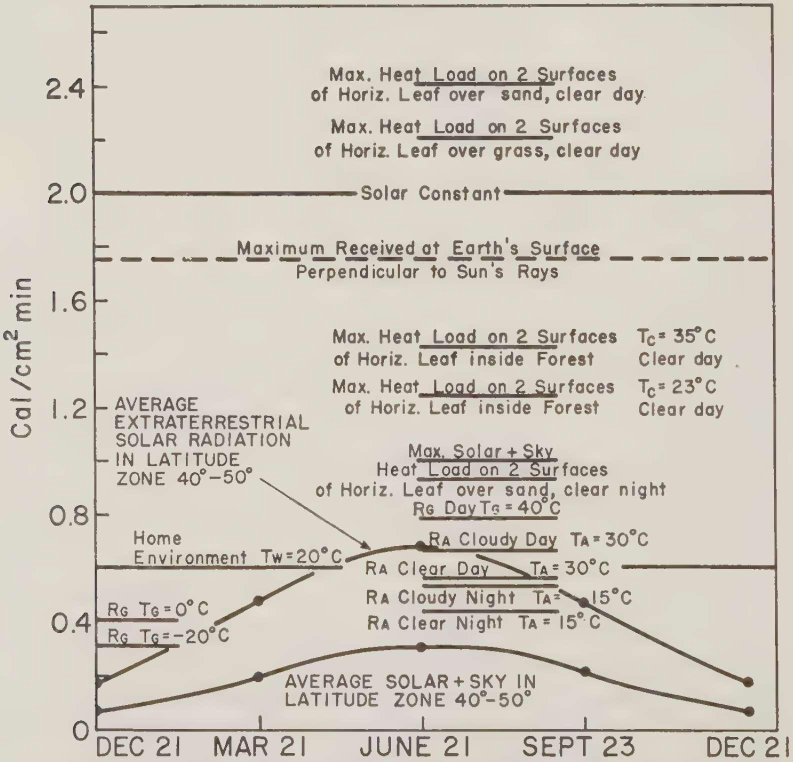


FIGURE 10.—The estimated maximum heat load received on the upper and lower surfaces of a horizontal leaf in various environments. The temperature of the forest canopy is T_C ; temperature of the air, T_A ; temperature of the ground, T_G ; temperature of the walls of a room, T_W ; radiation from the ground, R_G ; and radiation from the atmosphere, R_A .

radiation on a horizontal surface for the latitude zone 40° to 50°. The maximum value for July at latitudes 40° to 50° for solar plus skylight is given as a horizontal line at the value 1.0 cal per cm^2 per min. The maximum solar energy received at the earth's surface on top of the highest mountain on a plane perpendicular to the sun's rays would be approximately 1.75 cal per cm^2 per min and is shown by the dashed line at this level. Also for reference the value of 2.0 cal per cm^2 per min for the solar constant is also given.

On a clear night, with the surface air temperature at 15°C and with average humidity conditions, the downward stream of infrared thermal radiation from the atmosphere will be approximately 0.44 cal per cm^2 per min. If the night is overcast and the air temperature is 15°C, the downward stream of radiant flux is greater than on a clear night and is now approximately 0.53 cal per cm^2 per min. These two conditions are shown in Figure 10 as the lowest two horizontal lines. The next horizontal line represents the downward stream of atmospheric infrared radiation from a clear daytime sky at an air temperature of 30°C. The value here is 0.56 cal per cm^2 per min. Actually for calculating the stream of infrared thermal radiation from the atmosphere, it makes no difference whether it is day or night since the atmospheric long-wave radiation term does not include scattered or direct sunlight, but only the radiation thermally emitted by the atmospheric gases, water vapor, carbon dioxide, and ozone. Therefore the important parameters, which may vary with time and place, are the air temperature and the atmospheric water vapor content. If only the air temperature and relative humidity near the surface are known, then the atmospheric radiation calculated will be approximate to within about 15% or 20% of the true value. However, this accuracy is sufficient for the comparisons being made here. In order to obtain greater precision, it would be necessary to know the temperature profile and the humidity profile from the surface up to near the tropopause. Detailed calculations of this type will be described in the following lecture. If the day is overcast, the downward stream of infrared atmospheric radiation is stronger due to the fact that the clouds act as black bodies emitting at the temperature of their bases. An example of such a situation for an air temperature of 30°C near the surface is given by the fourth horizontal line from the bottom in Figure 10 with a value of 0.66 cal per cm^2 per min.

Also of importance to vegetation and animals, in addition to the downward stream of atmospheric radiation, is the upward stream of infrared thermal radiation from the ground. The ground radiates very much like a black body, and for a surface temperature of 40°C the upward stream of infrared radiation would amount to 0.79 cal per cm^2 per min and is shown by the horizontal line in Figure 10 marked by R_G. Higher ground surface temperatures can prevail under many circumstances during the daytime, and the radiated flux will be higher according to the fourth power of the absolute temperature. If the ground surface temperature was 60°C, the radiated flux would be 1.0 cal per cm^2 per min, about equal to the total sunlight received on a horizontal surface in July at latitudes 40° and 50° N. The reason for presenting Figure 10, as given here, is to show as emphatically as possible the fact that the long-wave infrared thermal radiation is indeed competitive with the solar radiation in terms of total energy. This fact becomes even more striking when one integrates the values over a twenty-four hour period. The total thermal energy radiated by a cloudy sky with surface air temperature 30°C over a twenty-four hour period would be 86 cal per cm^2, while the total amount of solar energy received on a clear day in June or July would amount to about 560 cal per cm^2. At all other times of the year the total solar energy will be less than this, and for cloudy days it will be considerably less. The upward stream of thermal radiation from the ground surface will also total about 860 cal per cm^2 if the surface temperature averages 20°C over a twenty-four hour period. Almost never will the thermal radiation stream, downward or upward, contribute less energy over a twenty-four hour period to a horizontal surface than will the solar radiation. The solar energy is of course important because it irradiates plants with the wavelengths of importance to photosynthesis and furthermore contributes energy to all of the dynamic and life-giving processes on the planet earth. But if the thermal radiation from the ground and atmosphere were not present, then life on this planet would not be possible either, for during the nighttime everything would get desperately cold. Such is the condition on the moon where there is no atmosphere—blazing hot by day and way below freezing at night. The point that should be made is that all components of the energy regime are important and to consider one, such as the solar, without considering the others is to study only a fragment of the total picture.

It is interesting to visualize the heat load that these streams of radiation place on the leaf of a plant. It is convenient to consider a horizontal leaf rather than one with some other orientation although other orientations could be worked out. If a horizontal leaf is in full sunlight over sand on a clear day, it will absorb a total radiant energy of 2.4 calories per two square centimeters of surfaces per minute. The two square centimeters involve one square centimeter of upward facing surface and one square centimeter of downward facing surface. The incident radiation on the upward facing surface was the sum of the total direct and scattered sunlight plus the downward stream of infrared thermal radiation from the atmosphere. The downward facing surface receives sunlight reflected from the ground plus the infrared thermal radiation emitted upward by the ground surface. The energy absorbed has already taken into account the reflectivity of the leaf surface. If the horizontal leaf is over a grass surface, the amount of sunlight reflected by the grass will be less than that reflected by the sand and the amount of energy absorbed by the leaf will be less. The maximum radiant energy absorbed by a leaf under these circumstances will be approximately 2.2 calories per two square centimeters of surfaces per minute. If one were to consider the average radiant heat load per surface it would amount to 1.2 and 1.1 cal per cm^2 per min, respectively, which is greater than the maximum amount of direct and scattered sunlight received on a horizontal surface and nearly four times the average value of direct and scattered sunlight. For a horizontal leaf inside a deciduous forest the downward stream of radiation is of quite different nature, in terms of its wavelength distribution and magnitude, than it is in the case of a leaf exposed directly to the sun and sky. The amount of solar flux will be greatly reduced and almost negligible in terms of energy contribution. The amount of infrared thermal energy from the atmosphere will also be negligible since it is assumed that the forest canopy above will cut off most of this radiation. The leaf will then receive thermal radiation emitted by the leaves of the canopy at some mean temperature for the canopy and also thermal radiation emitted from the ground surface. For a canopy temperature of 35°C the heat load will be 1.43 calories per two square centimeters of surfaces per minute, and for a canopy temperature of 23°C the heat absorbed will be 1.24 calories per two square centimeters of surfaces per minute. The values given in Figure 10 for the radiant energy absorbed by a horizontal leaf are maximum values for the conditions stated; for many actual situations somewhat smaller values will occur. Other situations in different environments should lead to values falling between and below the examples given here. The leaf itself must get rid of this radiant heat load, and there are three ways in which this can be done: by reradiation, by convection, and by transpiration. The leaf will radiate energy very nearly as a black body at the temperature of the leaf surface. It will reradiate at least half of the heat load. The convective transfer can be large or small depending upon the air temperature and the amount of wind. If the leaf is not subjected to much wind and if the air temperature is about the same as the leaf temperature, the convective term will be negligible. A substantial amount of energy will go into transpiration. If one works out these factors for day and for night conditions, one finds under many circumstances the leaf must continue to transpire on into the evening. Working out the heat load on a leaf and subtracting the reradiation component, one finds that the transpiration follows very closely to the net radiation as long as there is sufficient soil moisture.

For further comparison purposes, the radiant energy regime of a home environment with wall temperatures of 20°C is given. Any indoor space, such as a room, will act essentially like a black body cavity at the temperature of the walls. The infrared thermal radiation from the walls is an important component of indoor environments. One has experienced the discomfort of sitting in a room with the air temperature reasonably warm, say 72°F, and the walls cold, say 50°F. The lack of radiant heat from the walls can make one feel very chilled indeed. While mentioning indoor environments it is probably well to point out that similar problems pertain to environmental chambers, phytotrons, etc. Many experiments have been done in environmental chambers in which light flux has been carefully controlled and measured, and an utter disregard for wall temperature, thermal radiation, and convection processes has prevailed. For some experiments these oversights may not matter, but for other experiments these neglects could invalidate the conclusions.

Some environmental measurements made in Alaska by Stoll and Hardy (1955) are of particular interest. The measurements were made to assist with the problem of correctly clothing troops for survival in the arctic environment. Figure 11 is an example of conditions encountered at Nome, Alaska, under summer conditions. In order to evaluate the magnitude of the radiation environment, a thermal radiometer was used for measuring the flux of radiant energy coming from the sun, sky, clouds, or ground, and the strength of the radiation was expressed in terms of the emission by an equivalent black body at a definite temperature. In Figure 11, various radiant temperatures are given for sky, ground, and total environment. In addition, the energy from the sun is plotted and the total heat load on a human being in this environment is evaluated. The air temperature, T_A, was measured and found to remain constant within about 2°C over nearly a twelve hour period. The amount of cloud cover is plotted at the bottom of the figure. It will be noticed that the sky was clear from 7:00 A.M. to 8:30 A.M. Then the clouds began to form; the amount of direct solar radiation, R_o, received on a horizontal surface began to drop steeply; the equivalent environmental radiant temperature, T_W, fell steeply; and the heat load on a

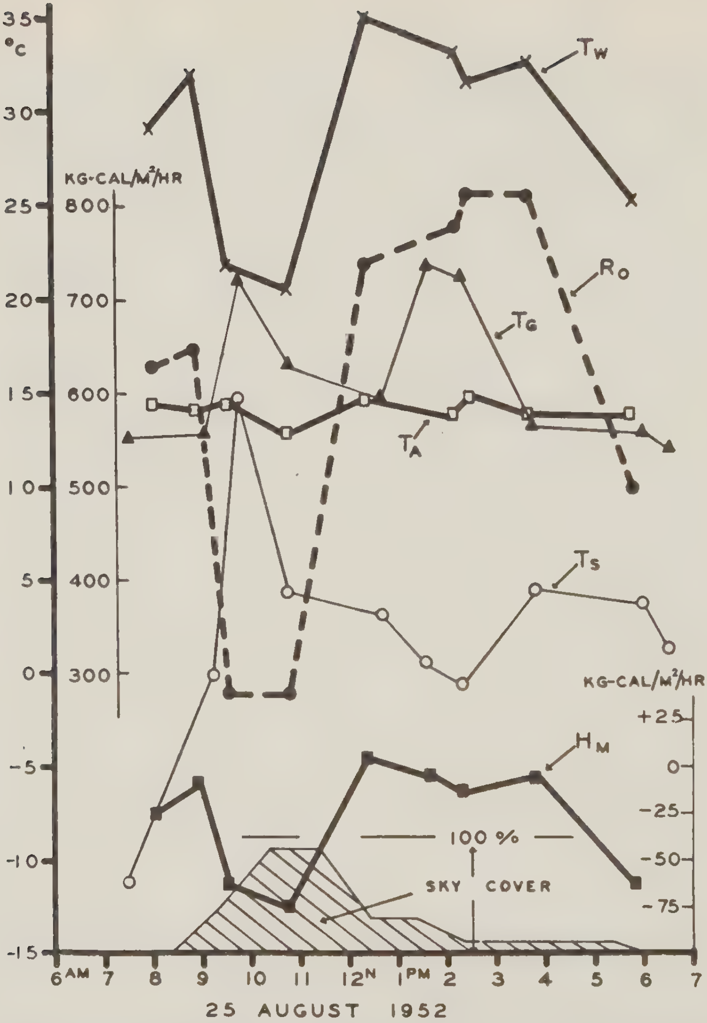


FIGURE 11.—Variation in solar radiation and environmental temperature and the resultant radiant heat load on man (August 25, 1952, Nome, Alaska). T_W is the radiant temperature of the environment; R_o, the direct solar and sky radiation; T_G, the radiant temperature of the ground; T_A, the air temperature near the surface; T_S, the radiant temperature of the sky; H_M, the radiant heat load on man; sky cover in % of total sky covered by clouds. (From Stoll and Hardy, 1955.)

man, H_M , dropped by nearly 75 kg-Cal per cm^2 per hr, a dramatic change in the heat load. Notice, however, the simultaneous steep increase in the radiant temperature of the sky because of the cloud cover. If it were not for this component of the heat load, i.e., the downward stream of infrared thermal radiation from the clouds, the drop in the heat load would have been even more severe. As the clouds thinned around noon the solar radiation began to penetrate and the radiant sky temperature dropped, but the radiant environmental temperature increased as did the heat load on man. Late in the day, as the sun went down and as the sky remained nearly clear, the radiant environmental temperatures began to drop and the heat load diminished correspondingly. This is a dramatic example of the sort of variation in the heat load which may occur daily for various types of environment throughout the world. The air temperature may remain nearly constant and the heat load may change tremendously. It is not sufficient to measure just the air temperature; in fact the air temperature may be very misleading as an environmental parameter by itself. It is absolutely essential to evaluate the flow of energy when considering the interaction of organisms with their environment. The response of the organism may have very little to do with the air temperature. There are relationships, of course, but they are not always as immediate and direct as one might imagine. Unfortunately, a great deal of wasted effort has been incurred over the last half century by taking detailed temperature measurements and then not making the proper use of them. Often expected or hoped-for correlations between temperature and some biological reaction, such as growth, did not show up and, in fact, in many cases it would have been surprising if they had. If one is to measure the temperature of part of the environment, such as the air, then it is essential to understand, in terms of the physics involved, the use to which such a measurement may be put. Specifically it may be of use to determine the convective heat transfer from the organism to the air and of some value in a determination of the thermal radiation emitted by the air or atmosphere. Air temperature by itself may be of only secondary importance in biological work unless properly utilized, although it is an extremely important parameter to the meteorologist.

In Alaskan winter, environmental temperatures can become frightfully cold when the air temperature is low and the air is clear and dry. Figure 12 contains a histogram of the radiant sky temperatures in the zenith. In addition it shows the average summer and the maximum and minimum winter tem-

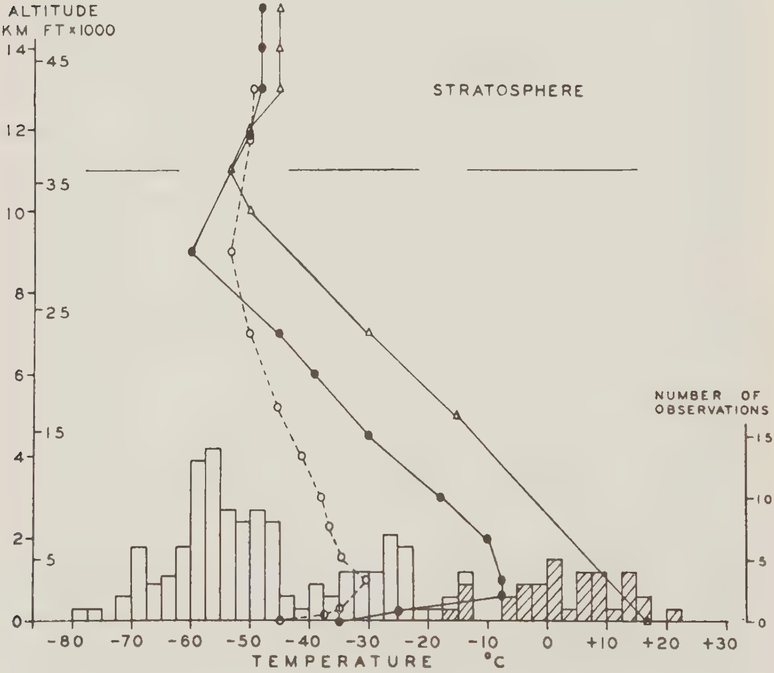


FIGURE 12.—Radiosonde air temperatures compared with radiant sky temperatures measured at the zenith. Air temperatures are plotted in °C on the abscissa, and altitude is plotted on the ordinate on the left. Solid line through open triangles represents an average of summer readings. Solid line on closed circles represents highest temperature recorded, and broken line on open circles represents the lowest temperature recorded during the winter experiments. The radiant sky temperatures are plotted in °C on the abscissa, and the number of observations is indicated on the ordinate on the right. (From Stoll and Hardy, 1955.)

perature profiles as a function of altitude as measured with a radiosonde. The important point to note here is that the radiant sky temperature on one occasion dropped to the extremely low value of nearly —80°C. This means that mighty little radiant heat was coming groundward from the clear, dry, cold sky. The lowest air temperature encountered near the ground was —45°C, and the lowest temperature in the air was —60°C at a height of approximately 28,000 feet. Thus we have once again an example of environmental conditions much more severe than the air temperature by itself would indicate—in fact, 35°C colder than the lowest air temperature near the surface. If arctic troops were clothed to withstand the air temperature conditions known to exist near the surface, they would rapidly freeze to death by losing heat through the dry, clear sky to the cosmic cold of outer space. The environmental temperatures in the arctic are usually 15°C to 20°C colder than the air temperature. Remember that the term "environmental temperature" is a fictitious quantity, being the temperature a black body would need to have in order to produce thermal radiation equal to that actually being received from the environment. The term "radiant sky temperature" has the same sort of meaning, i.e., the temperature a black body would have to assume in order to produce a stream of thermal radiation equal to the downward stream of radiation from the sky. If the sky were a completely transparent gas it would produce no thermal radiation, since it would not absorb any radiation, and the sky in this instance would have a radiant temperature of absolute zero. Fortunately this is not the case for the earth's atmosphere.

In Figure 13 can be seen one more example of the great excursions in the environmental radiant temperatures of the sky while the air temperature and the ground temperature remained nearly constant over a thirty hour period. The radiant sky temperature reached a minimum of —75°C when the air temperature was —38°C. When the sky became overcast and it began to snow, the radiant sky temperature rose 48°C to a value of —27°C, which was some 7° or 8°C above the air temperature at that time. Similar measurements reported by Gates

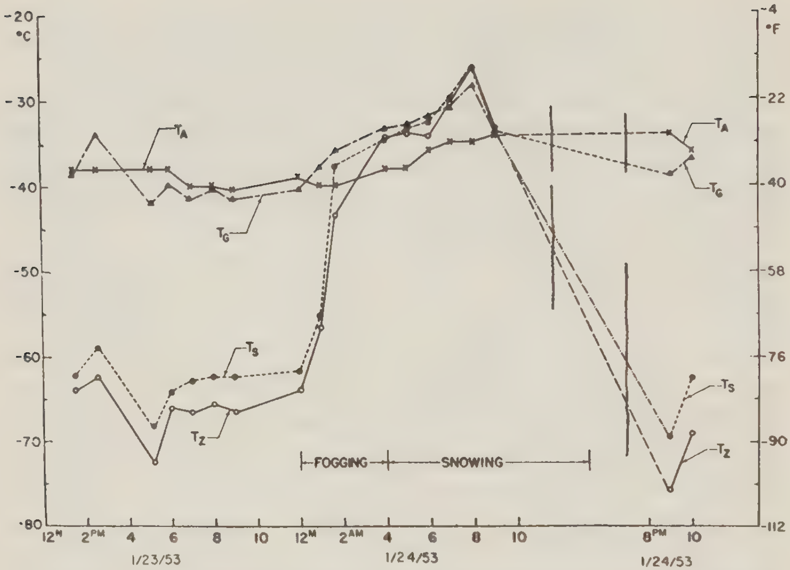


FIGURE 13.—Effect of fog and snow on environmental radiant temperatures. T_A is the air temperature; T_G, the radiant temperature of the ground; T_S, the radiant temperature of the sky; T_Z, the radiant temperature of the sky at zenith. (From Stoll and Hardy, 1955.)

(1961) were made in Yellowstone National Park during February, 1961, and are discussed in Chapter 5.

A different example of the behavior of sky temperature and air temperature is given in some measurements by Kelley, Bond, and Ittner (1957b) made in the Imperial Valley of California in August, 1957. The air temperature, measured at 42 inches above the ground, changed from 90°F to 103°F during the daylight hours 0700 to 1600. The radiant temperature of the north sky changed from 54°F to 82°F during the same period. These changes were entirely associated with the variations in cloud cover during the day.

Some interesting radiation measurements of a forest stand have been made in Bavaria by Baumgartner, who was studying the heat budget of a pine forest. The hourly totals of the net radiation are shown in Figure 14. Observations were made with a nonventilated net radiometer at 10 meters height above the tree tops, 4.1 meters height in the crown of the trees, and 2.1 meters height at tree trunk level. The height of the pine trees was approximately 6.0 meters. At the 10 meter level above the tree tops the net radiation is about as expected, showing strong insolation in the daytime and fairly strong cooling at night. In the crown of the trees the amount of net positive flux is somewhat reduced because of the screening effect of the tree tops for low solar angles. The asymmetry of the curve is due to the geometry of the tree tops. At the tree trunk level there is nearly a two hour lag in the time of occurrence of positive net flux in the morning due to the screening from the sun. The maximum insolation reaching the ground is only 20% of that reaching the tree tops, and nega-

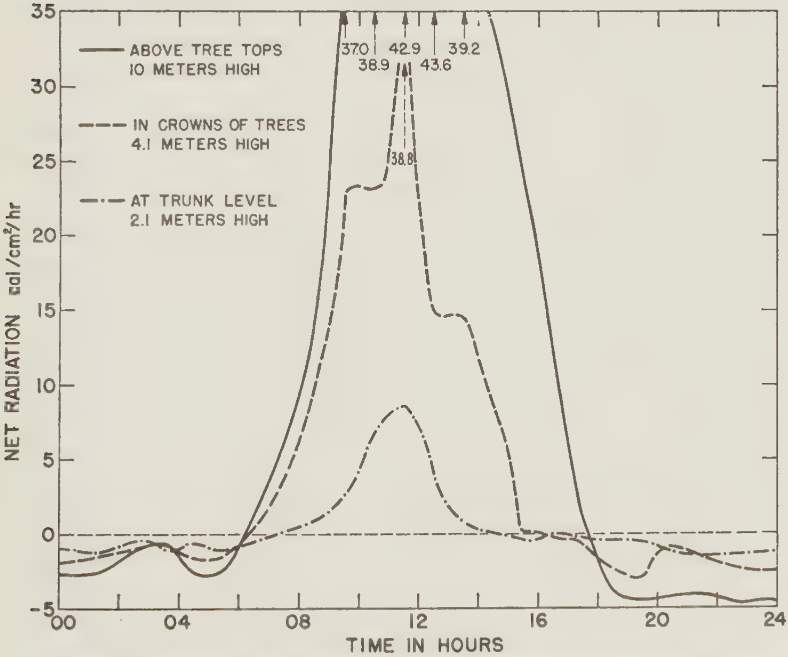


Figure 14.—Hourly totals of the net radiation in a stand of pine trees in Bavaria. (From Baumgartner, 1956a.)

tive net flux is re-established more than three hours before sunset. It is interesting to compute the total accumulated value of the net flux at the three levels. The accumulated negative net flux during the night has been completely compensated for by the positive net flux of the daytime by 0830 at 10 meters height, by 0900 at 4.1 meters, and not until 1030 at 2.1 meters. The total accrued energy during a twenty-four hour period was in the ratio of 100:40:5 for the space above the trees, to the space in the crowns, to the space at the tree trunk level, and the number of hours of positive net flux were in the ratio of 8:6:4. Over a three day period of observation, the air at the 10 meter height received a positive accumulation of 850 cal per cm^2, at the 4.1 meter level 350 cal per cm^2, and in the 2.1 meter region 35 cal per cm^2. Ordinarily, such energy would raise the temperature of the air above the tree tops several hundred degrees; however, the temperature actually rose only 2.7°C. In the crown of the trees it rose 3.0°C and at the trunk level 0.5°C. Clearly much of the excess energy must have been disposed of through turbulent exchange and evaporation. The interior of a forest may be a very gentle place indeed compared to the harsh conditions of more exposed environments. The important lesson to be learned by this example is that the usual meteorological parameters could not have given an accurate description of the microclimatology of this stand of pines. The energy and moisture budget must be dealt with quantitatively.

In a most interesting paper, Miller (1956) presented a detailed study of the heat transfer in the needles and crowns of a stand of lodgepole pine. This study was an attempt to explain the unusually warm air (often 60°F) encountered above snow-covered ground in and around the stands of pines in the Sierra Nevada mountains. Miller assumed the incoming solar radiation incident on a horizontal surface to be 90 cal per cm^2 per hr. By virtue of the vertical nature of the crowns and the spacing between them, he estimated that a total of 154 cal per cm^2 per hr is absorbed in the crowns, 15 cal per cm^2 per hr are transmitted through the crowns to the ground, 18 cal per cm^2 per hr are reradiated upward to the sky, 11 cal per cm^2 per hr are reradiated downward, 3 cal per cm^2 per hr go into photosynthesis, 2 cal per cm^2 per hr are stored in the wood, 80 cal per cm^2 per hr are transferred to the air through convection, and 40 cal per cm^2 per hr are transferred to the air by transpiration. Therefore, on a typical spring day, about 120 cal per cm^2 per hr of heat are supplied to the air by the needles of the pine forest. Miller wrote, "as the sun heats the trees, the level of highest temperature migrates from 50 feet downward to 30 feet and lower. For several hours this middle level is warmer than the air either above or below it, and heat flows both upward and downward from it." There is no doubt that similar phenomena should occur in coniferous forests throughout the world.

ANIMAL RESPONSE TO ENERGY ENVIRONMENTS

Having learned something concerning the energy factors of various environments, it is of interest to consider the response of plants and animals to their environment. Various organisms will behave differently with respect to their mode of energy exchange in any given environment. Some will lose most of their heat by radiation, some mostly through convection, and some mostly through evaporation, while some will distribute the loss among these mechanisms. Therefore, animals might be classified according to whether they are radiators, convectors, or evaporators. Unfortunately most of the research to date has used temperature as a frame of reference and rarely in the literature is there any mention of the energy regime.

Large insects such as the locust (Schistocerca gregaria), the cockroach (Periplaeta americana), and the grasshopper (Gastrimargus transversus) are able to maintain their body temperature below the ambient temperature of the air when there exists a substantial heat load. The evidence seems to be that they can do this better in dry air than in moist air. The sun shining on insects definitely increases their body temperature and sometimes at a fairly high rate. A small animal is at a distinct disadvantage in keeping its body temperature down by means of evaporation since the water available soon becomes exhausted. A cockroach loses about 80% of its energy loss through convection and about 20% through radiation. Work done by Parry (1951) in England has shown that insects compare very favorably with inanimate dry objects of the same size and shape as far as the mechanism of energy exchange is concerned. This is consistent with the fact that the factors metabolism and evaporation play an insignificant role in the energy exchange of insects. Parry carried out a series of experiments on inanimate objects to determine the details of the heat transfer mechanism. This is an extremely important philosophy since if one cannot understand how a simple cylinder or a flat plate exchanges energy in a given environment, then it is absurd to attempt to understand the heat exchange mechanism for an insect. Insects appear to be especially adapted for adjusting to their environment and surviving under extreme conditions. Some insects, such as locusts, are able to sun themselves and orient their bodies favorably for catching heat. Insects of arid zones have special adaptations such as long legs to keep their bodies up off of the hot sands and a partial or total loss of the power of flight. In Algeria and Tunisia nearly half of the arthropods are wingless. Many Tenebrionidae and Carabidae have fused wings. The discussion of the energy regime of the desert of Aswan, Egypt, showed that where there was a lack of moisture and a strong heat load, a windy, turbulent condition resulted. The deserts are characterized as windy environments. Where the winds are frequently strong the insects apparently need to remain earth bound much of the time. In other parts of the world where the heat load may be high but the winds are low, insects will use their mobility to fly through the air, thereby generating heat loss through forced convection. In the grass near the ground, in our own yards during the summer, conditions are essentially tropical, hot and moist, and insects often must use their mobility to climb up a stalk of grass to get in the wind so that the convective heat transfer will cool them off. With regard to insects in cold environments there seems to be little doubt that a few genera can survive very low temperatures and the formation of some ice within their tissues; however, complete freezing would appear to result in death in most cases. Some insects apparently generate antifreeze in their body plasma in the form of glycerol and survive subfreezing conditions as low as —30°C. These insects generate the antifreeze during the autumn and winter and usually perish when exposed to the same low temperatures during the spring and summer.

Amphibia are interesting in that they would be classified as evaporators. They act like a wet-bulb thermometer when they emerge into the open air with even the slightest wind. The skin of amphibia is more transparent to sunlight than is the outer layer of many animals and hence they heat up more rapidly. Amphibia frequent moist and shady places in order to protect themselves against the danger of death through desiccation.

Reptiles gain and lose heat by radiation and convection but not through evaporation. The mobility factor is critical for reptiles. They are able to seek out warm exposures to the sun, change their body orientation if they become too warm or too cool, or retire to the shade if necessary. Some lizards can even modulate their skin pigmentation according to their need for capturing solar radiation. It is always interesting to speculate on the behavioral mannerisms of the giant reptiles. It is clear they would be late to rise and to retire because of their large body mass and heat capacity. The giant reptile would be very lethargic during the morning hours but would gradually warm up as the sun heated the body. It might be almost noon before he generated much pep. Then in the evening, warmed by the stored solar energy, he would forage for food, gradually lose heat by reradiation, and finally settle down about midnight.

The role of evaporation in the energy exchange of various animals is extremely different when expressed as the percentage of body weight lost per day when exposed to warm, dry environments. The python and tortoise lose about 0.1% to 0.3% and therefore do not exhibit any evaporative control of their body temperatures. The meal worm loses about 1%

of his body weight per day, the cockroach as high as 4%, and the wood louse 14%. The salamander would perish very quickly in dry air, losing moisture at such a rate as to lose the equivalent of 950% per day.

Birds are magnificently constructed to conserve heat. The body feathers contain myriads of air spaces, thereby producing excellent insulation. The extremities of birds are poorly insulated as they consist of bone and tendon with little blood flow and can function at temperatures much lower than muscles, and heat loss from them must be small. Birds must lose a significant proportion of their body heat through radiation and some by evaporation through breathing.

Mammals, along with birds, are endowed with the ability to maintain a nearly constant body temperature. Various mammals must differ considerably in their heat transfer mechanisms. Man, cats, and dogs, when the heat load becomes too great, open their mouths, pant, and salivate. Pigs and rabbits pant less and do not open their mouths. Rats and mice accelerate their breathing very little and do not open their mouths when subjected to a heavy heat load. A sheep loses approximately 48% of its energy loss through radiation, 38% by convection, and 14% by evaporation. A sheep in the open during the summer, exposed to the full blazing sun, is subjected to a pretty severe heat load. Mcfarlane (1958) in Australia measured the temperature in the wool of a sheep's back to be as high as 87°C or nearly 190°F. Livestock usually seek out shade under such circumstances, and by standing on the north side of a tree (for the northern hemisphere) they not only will be shaded from the sun, but will radiate heat and, in addition, receive less radiation in return from the "cold spot" in the sky which exists in the northern sector at 90° from the sun. This knowledge of the "cold spot" in the sky has been very useful for properly designing shelters for animals in hot summer climates.

Many detailed studies have been made concerning the heat transfer process with regard to man. Among the classic works in this regard are those by Winslow, Herrington, and Gagge (1936a,b, 1938a,b), Gagge (1936, 1940), Gagge, Winslow, and Herrington (1938), and Winslow, Gagge, and Herrington (1940). These investigators invented the measurement technique of partitional calorimetry and by this means sorted out the various terms of the energy budget equation as applied to man. They found the unclothed man to lose 47% of his energy loss through radiation and 53% through convection. A clothed man, dressed lightly in underwear and a business suit with a vest, loses 55% through radiation and 45% through convection. The absolute total loss of energy is, of course, less for the clothed subject than for the unclothed, but the mechanisms for losing energy are in the proportions given here. Clothing constitutes insulation which tends to retard the flow of air over the surface, thereby reducing the loss by convection compared to the loss by radiation. The type of measurements which have been made on man are absolutely fundamental and should now be made on many other organisms of the animal kingdom. It is important to establish for each organism the fundamental constants which define the flow of energy between the organism and its environment. The constants are the radiation emissivity, the conduction coefficient, the convection coefficient, the evaporative properties, and the metabolism and photosynthesis rates. Modern science has now matured to the stage where environmental investigations can be done in rigorous fashion. Modern biology, including ecology, requires a detailed knowledge of the physical world as well as of the biological world. The knowledge and skills are available from physics, chemistry, mathematics, and biology, but the biologist must avail himself of the training and knowledge available in these other disciplines. A detailed understanding of the mechanisms of interaction of the organism with its environment will lead to dramatic advances in our understanding of the distribution and propagation of the species on the planet earth, will give quantitative information of importance to genetics and origins, and will, for the first time, place the subjects of paleobotany and paleontology on firmly analytical bases. For modern biology there is no choice, for we have the ability, the knowledge, and the techniques, providing the desire exists, to properly integrate the sciences and to truly consider the organic world as an integral and vital part of the physical world.

Ecology—the interaction of plants and animals with their environment—is indeed an exciting and rewarding subject. However, it is an extremely difficult discipline if properly undertaken. Progress in the subject has been steady, if not always spectacular, but now a new era of immense advancement may begin if biologists avail themselves of the techniques and ideas of physics, chemistry, and mathematics.

Chapter 2

Solar and Thermal Radiation

INTRODUCTION

The purpose of this and succeeding chapters is to give a more detailed technical discussion of the physical principles described in the first chapter for the use of the biologist. The sun, directly or indirectly, is the source of most of the energy derived by the planet earth. The amount of solar radiation reaching the earth, the rotation period of the earth on its axis, and the composition of the atmosphere all combine to make the evolution of life on the planet possible. For this reason a great deal of effort has been centered on the study of solar radiation, its distribution over the earth, and its utilization. Detailed material on this subject will not be reproduced here; however, references will be given for finding the appropriate information.

SOLAR RADIATION

The best evaluation of the solar constant—the flux of energy outside the atmosphere on a surface perpendicular to the sun's rays at the mean distance of the earth from the sun—is presented by Johnson (1954), who arrived at the value 2.0 cal per cm^2 per min. The wavelength distribution of the solar energy is shown in Figure 3. Even outside of the earth's atmosphere, before depletion by the absorbing constituents, the energy curve is not a simple, smooth, consistent envelope, but has structure to it with spikes of energy appearing on the short wavelength, ultraviolet side. The general features of the solar energy curve can be approximated by a 6000°K black body radiator, but in places the deviations from this curve are large indeed. When the energy distribution is plotted against wavelength, the maximum monochromatic intensity of the sun would appear to occur at 0.47 microns in the green of the visible spectrum. This is an illusion because of the method of plotting. It is more correct to plot the energy versus the frequency, in which case the area under any section of the curve is directly proportional to the energy. The reason for this is that energy in the electromagnetic spectrum is equivalent to $\hbar\nu$, where $\hbar$ is Planck's constant and ν is the frequency of the radiation. The frequency may be given in the customary units of vibrations per second or it may be given in other units proportional to this. The fundamental relationship for relating wavelength to frequency is: $\lambda\nu = c$, where c is the velocity of light in cm per sec, λ is the wavelength in cm, and ν is the frequency in vibrations per second. However, since c is a constant it is possible to write $\lambda\nu = 1$ with λ in cm and ν in reciprocal centimeters or cm^{-1}. When energy is plotted versus frequency as shown in Figure 15, the true energy peak is shifted to just beyond 10,000 Å or 1.0 micron. The median value occurs at 7100 Å in the near infrared just beyond the visible. Twenty-nine per cent of the energy lies at wavelengths greater than 1.0 micron and 50% beyond the red of the visible. The distribution is somewhat shifted after the solar radiation has traversed the earth's atmosphere to the surface. The spectrum is continuous, extending throughout the infrared to 1000 microns where begins the portion of the spectrum termed the microwave—the shortest of the radio frequencies. The spectrum from the sun continues, and much longer radio wavelengths are emitted from sources deep within the solar atmosphere at temperatures as high as 100,000°K. Thus the statement so often made in textbooks—that the solar spectrum ends at 3.0 microns or thereabouts—is quite false. For the extraterrestrial radiation the spectrum continues almost indefinitely, and several segments of the spectrum well beyond 3.0 microns reach the earth's surface. The energy becomes less and less toward longer wavelengths, and the contribution of these segments on the tail of the curve to the total energy received is almost

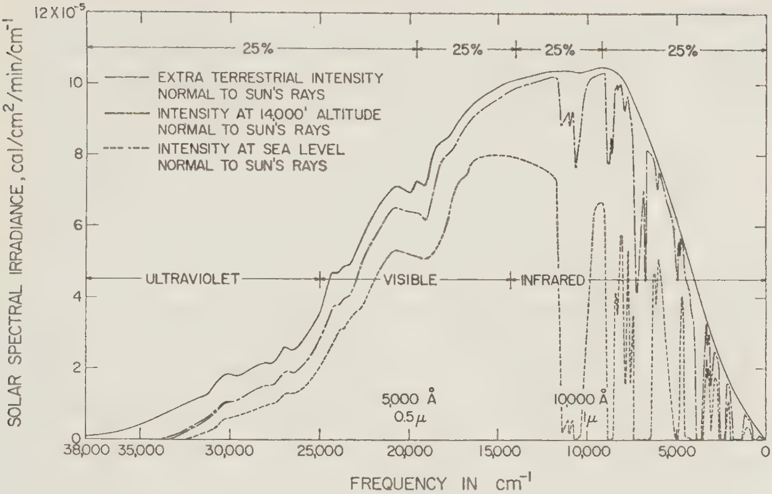


FIGURE 15.—The solar spectrum at the mean solar distance from the earth as received outside of the earth's atmosphere on a surface perpendicular to the solar rays. The intensity scale is based on a solar constant of 2.00 cal per cm^2 per min per cm^{-1}, which would represent the area under the upper curve. The intensity of direct beam sunlight received on surfaces perpendicular to the solar rays at an altitude of 14,000 feet and at sea level for a zenith sun is also shown. The attenuation at 14,000 feet is based on ozone amounts of 2.00 and 3.00 mm which results in the splitting of the curve at the high frequency end. Two amounts of total precipitable water vapor content are used: 1.0 mm and 25.0 mm, representing the extremes to be encountered.

negligible. The total solar energy arriving at the earth is almost constant with time; however, tremendous increases occur in the very short wavelength, very high frequency, ultraviolet and X-ray regions, coincident with the appearance of a flare or disturbance of the solar corona. Although these impulses of ionizing radiations may change by many times their ambient values, the contribution of these changes to the total energy from the sun is negligible. It is still very likely that the bursts of short wavelength radiation, occurring at times of solar activity, are of considerable importance biologically in that the earth is flooded with highly penetrating X rays, which in turn could produce genetic mutations in organisms if these X rays were to reach the surface. Normally all of these X rays have been stopped high in the stratosphere at balloon heights around 100,000 feet above the surface. Although the most intense X rays have been observed lower in the atmosphere, they have never been observed reaching the surface. The daily extraterrestrial, undepleted insolation at any position of the earth's surface is dependent only upon the latitude of the station and the time of the year. These daily totals are shown in Figure 2. The maximum value of daily total extraterrestrial insolation occurs at the summer pole having twenty-four hours of sunlight and amounts to about 1060 cal per cm^2 per day. The average undepleted insolation over the sunlit hemisphere is about 0.973 cal per cm^2 per min. The long-term mean value of extraterrestrial insolation averaged over the globe for an entire year is 0.485 cal per cm^2 per min, and this is the value used for computing the numbers shown in Figure 1. It is important to appreciate the extent to which some regions of the earth, such as the polar latitudes, are subjected to a strong annual pulsation of solar heating and cooling, while equatorial regions exist in a nearly uniform diurnal flood of sunlight so that the daily totals vary only slightly from winter to summer.

The stream of solar radiation is depleted by scattering and absorption on passing through the earth's atmosphere to the surface. The amount of dust clouds and absorbing constituents in the atmosphere vary greatly over the surface and from day to day. For this reason the insolation at a given location on the surface is an ever-changing quantity dependent upon local atmospheric conditions. The solar radiation passing through the atmosphere is affected by the absorption and scattering processes according to the following exponential law:

$$I_\lambda = I_{\lambda o}\, e^{-(k_\lambda + s_\lambda)\, u \sec \theta} \tag{2.1}$$

where

I_λ = monochromatic intensity of the solar beam at the earth's surface.

$I_{\lambda o}$ = monochromatic intensity of the solar beam outside of the earth's atmosphere.
k_λ = absorption coefficient at the wavelength λ to $\lambda + d\lambda$.
s_λ = scattering coefficient at the wavelength λ to $\lambda + d\lambda$.
u = optical depth of the absorbing or scattering material in grams per cm^2 of vertical column.
θ = zenith angle of the sun.

The absorption and scattering coefficients are both complex functions of the wavelength. The absorption coefficient varies widely with wavelength due to the fact that each absorption band consists of hundreds of individual, narrow absorption lines of various strengths which are often temperature and pressure dependent. Values of the absorption coefficients for the atmosphere have been determined in papers by Dunkelman and Scolnik (1959) and by Gates (1960). The amount of absorption of sunlight produced by atmospheric carbon dioxide and oxygen in the atmosphere should be nearly uniform over the surface, except for the effect of the changing altitude of the surface above sea level, since these gases are uniformly mixed in the atmosphere. The absorption produced by water vapor will be strongly variable over the globe due to the heterogeneous distribution of water vapor in the troposphere. The variations in absorption due to ozone will be small, although they will affect the ultraviolet cutoff which changes with altitude. It is well known that more ultraviolet radiation will penetrate to the surface of mountain tops than to lower elevations (see Stair, 1952). Bener (1960) presents recent research concerning the ultraviolet radiation received from the sun and sky on cloudless days and includes references to all earlier work. In principle it would be possible, knowing the amount of water vapor and the density of the atmosphere above a station, to compute from Equation 2.1, wavelength for wavelength, the transmission of solar radiation and then to integrate over the wavelength span to get the full energy flux received at the surface. This is far too troublesome to be worthwhile, and our information of the state of the atmosphere above a given station is not sufficiently accurate to make this a reasonable mode of operation. The scattering coefficient, s_λ, is also a complex function of the wavelength. Scattering by air molecules, nitrogen, oxygen, etc., depends directly upon the index of refraction and the density of air, and inversely upon the fourth power of the wavelength. This is the well-known Rayleigh scattering law. Usually larger dust particles are also in suspension in the atmosphere and produce an entirely different form of scattering. Ångstrom (1951) has shown empirically that the scattering coefficient for dust should be of the form

$$s_\lambda = \beta\lambda^{-\gamma} \tag{2.2}$$

where β is proportional to the particle density and γ is a parameter which decreases as the ratio of particle diameter to wavelength increases. So with larger size particles in suspension the value of γ drops toward unity. Average atmospheric conditions give the value 1.3 and in dust storms the value 0.5. For fog droplets, $\gamma = 0$, a condition referred to as diffuse reflection, in which the scattering is independent of wavelength.

It is usually best to separate the water vapor absorption from the dry, clear sky absorption and also from the scattering depletion due to dust. Brooks (1959) has done this for the direct beam of solar radiation expressed in BTU per ft² per min. Rewriting this for the flux of direct sunlight in cal per cm² per min received on a surface perpendicular to the sun's rays, one gets:

$$Q_p = \frac{2.0}{r^2} \exp\left[-0.089(pm/1013)^{0.75} - 0.174(wm/20)^{0.6} - 0.083(dm)^{0.9}\right] \tag{2.3}$$

where

Q_p = flux of direct sunlight on a surface perpendicular to the sun's rays in cal per cm² per min.
p = the air pressure in millibars.
w = the total precipitable water vapor of the at-

mosphere in the zenith direction in mm. (Precipitable water vapor is the thickness of liquid water which would be formed if all of the water in the atmosphere were condensed into a liquid film at the surface.)

d = the concentration of haze and dust particles in number of particles per cc.

m = the air mass by which is meant the ratio of the slant path length of the sun's rays through the atmosphere to the zenith path length or approximately the secant of the zenith angle of the sun. The zenith angle of the sun will be given by Equation 2.9.

r = the radius vector of the earth, which is the distance from the earth to the sun expressed as a fraction of the semimajor axis of the earth's orbit.

The value of w will run about 1 mm for a very cold, dry atmosphere in the winter to as high as 30 mm for warm, moist summer conditions, as shown by Gates (1956). These values can usually be obtained from the standard U.S. Weather Bureau radiosonde soundings by integrating the moisture profile with height. The values for d may vary from 0.2 to as high as 3.0 particles per cc. Usually for clear days d will vary from 0.6 to 1.0 and for polluted atmospheres in the vicinity of cities from 1.4 to 2.0. The value for d is likely to be the most difficult to obtain for a given situation. The air mass values will run from very large, 15 or so, early in the morning, to as low as 1.0 if the sun is in the zenith. Usually values of practical interest will range from 1.4 to 5.0 corresponding to solar zenith angles of 45° to 78°. The formula is more useful for computing the dust content of the atmosphere when one measures Q_p, p, m, and w, all of which are readily measureable.

Of more practical value are the large numbers of measurements of the daily insolation at the earth's surface made at many stations distributed over the globe. The meteorologist may be interested in large zonal accumulations of this energy, and these are given in a number of publications, notably Haurwitz (1941). The *Smithsonian Meteorological Tables* (Sixth Revised Edition, 1951) gives a chart of the undepleted solar radiation at the top of the atmosphere and numerous graphs and tables for computing the depleted insolation strength at the surface. A series of articles by Hand (1947, 1949, 1950, 1953, 1954) gives the values of the daily totals for direct and sky scattered sunlight on a horizontal surface from numerous stations throughout the United States. These articles describe how the average annual insolation received at the surface depends upon such factors as cloudiness, fog occurrence, and location near bodies of water, on windward side of mountains, or in a dusty, contaminated region. The annual receipt of solar energy in the Southwest is twice that received in the Great Lakes region. The Northwest coast and the Mississippi Delta regions rank next to the Great Lakes region in the lowest amount of annual solar energy received in the United States due to cloudiness factors. Liu and Jordan (1960) have presented procedures for evaluating the intensity of diffuse radiation on clear days and long-term average hourly and daily sums of diffuse radiation from a knowledge of the total (direct plus diffuse) radiation received on a horizontal surface.

Fritz and MacDonald (1949) have given maps of the average solar energy received over the United States for each month of the year and have worked out a linear regression equation relating the amount of energy received at the surface on cloudy days with the amount received on clear days and the number of hours of sunshine. The corresponding equation is as follows:

$$Q = Q_o(0.35 + 0.61\,S) \tag{2.4}$$

where Q is the amount received on a horizontal plane at the surface on any day, Q_o is the amount received on clear days, and S is the number of hours of sunshine instrumentally recorded divided by the number of hours of possible sunshine during the day. Q_o, the solar flux received on a horizontal surface, is related to Q_p, the solar flux received by a surface perpendicular to the sun's rays, by $Q_o = Q_p \cos z$, where z is the zenith angle of the sun.

Black (1956), Budyko, Berliand, and Zubenok (1954), and Burdecki (1956) have given analyses of the relationship between the amount of sunlight received and the cloudiness conditions. By means of a statistical analysis of solar data (direct sun plus sky), from 88 stations distributed over the globe, Black deduced the following empirical formula:

$$Q = Q_o' \,(0.803 - 0.340C - 0.458C^2) \qquad (2.5)$$

where Q_o' is the extraterrestrial solar radiation in cal per cm^2 per day, such as given in Figure 2, and C is the cloud amount in fraction of the sky covered, usually expressed in tenths. This formula assumes that the maximum amount of sunlight received at the surface on a cloudless day is only 0.803 of the amount received on a similarly oriented surface outside of the atmosphere. This is not a completely valid assumption since atmospheric conditions such as dust concentration may be very different on various cloudless days. Thus when the amount of cloudiness is known, the amount of sunlight received at the surface can be estimated by this formula.

A more accurate and generally more useful formula is one derived by Ångstrom (1922) and modified by Savinov (1933):

$$Q = Q_o[1 - (1 - k)C] \qquad (2.6)$$

where Q_o is, as in Equation 2.4, the total amount of solar energy received by a horizontal surface at the ground on clear days, C is the amount of cloud cover in tenths, and k is a parameter which is a function of latitude having the values 0.47 at 70°, 0.39 at 60°, 0.32 at 50°, 0.30 at 40°, 0.29 at 20°, and 0.32 at 0°. It will be seen, however, that k actually represents the ratio of the total sunlight received when there is complete overcast, $C = 1$, to the total sunlight received through clear skies. The values of k given here are only broadly representative of latitude belts each of which are characterized by certain cloud types. The values listed in Table II give k as a function of cloud type and solar angle or air mass for completely overcast conditions. Therefore, for a particular cloud type in the sky, one could select more accurately the value of k to use in Equation 2.6. One could compute values for various times of the day, therefore for different air mass values, apply Equation 2.6, and then integrate the values to get the total insolation for the entire day.

Haurwitz (1948) has made an extensive study of the relationship between the insolation received through an overcast sky and the type of cloud cover. He fitted the following empirical relationship to a large number of observations which had been made at Blue Hill Observatory near Boston, Massachusetts:

$$Q = Q_o(a/94.4)\, e^{-(b - 0.059)m} \tag{2.7}$$

The term m represents the air mass or the ratio of the slant path length for the direct solar radiation through the atmosphere to the path length in the zenith direction. The terms a and b are parameters dependent upon the type of cloud cover, and their values for each type are listed in Table I. In Table II are given the values of the ratio Q/Q_o which is the ratio of the insolation received through overcast sky to the insolation received through clear sky.

Formulas of the above types are only approximations at best, but are nevertheless very useful. Except for direct measurement, which is often difficult, the only other method for estimating the amount of solar energy incident at the surface is through the use of formulas such as these. Furthermore, these formulas will not give the instantaneous values of the total solar irradiation, but will give reasonably good

TABLE I

Cloud Cover Parameters for Solar Insolation as Influenced by Various Cloud Types (Ci, cirrus; Cs, cirrostratus; Ac, altocumulus; As, altostratus; Sc, stratocumulus; St, stratus; Ns, nimbostratus)

	Ci	Cs	Ac	As	Sc	St	Ns	Fog
a	82.2	87.1	52.5	39.0	34.7	23.8	11.2	15.4
b	0.079	0.148	0.112	0.063	0.104	0.159	−0.167	0.028

TABLE II

RATIO OF INSOLATION WITH OVERCAST SKY TO INSOLATION WITH CLOUDLESS SKY FOR VARIOUS CLOUD TYPES (Ci, CIRRUS; Cs, CIRROSTRATUS; Ac, ALTOCUMULUS; As, ALTOSTRATUS; Sc, STRATOCUMULUS; St, STRATUS; Ns, NIMBOSTRATUS)

Air Mass	Ci	Cs	Ac	As	Sc	St	Ns	Fog
1.1	0.85	0.84	0.52	0.41	0.35	0.25	0.15	0.17
1.5	0.84	0.84	0.51	0.41	0.34	0.25	0.17	0.17
2.0	0.84	0.78	0.50	0.41	0.34	0.25	0.19	0.17
2.5	0.83	0.74	0.49	0.41	0.33	0.25	0.21	0.18
3.0	0.82	0.71	0.47	0.41	0.32	0.24	0.25	0.18
3.5	0.81	0.68	0.46	0.41	0.31	0.24		0.18
4.0	0.80	0.65	0.45	0.41	0.31			0.18
4.5					0.30			0.19
5.0					0.29			0.19

estimates of the total amount of energy received during the day.

Klein (1948) has given an excellent discussion of the role of various factors affecting insolation and among others discusses the effect of altitude. According to Klein, the following factors must be considered as altitude dependent: (*a*) "the air mass at an elevation of z meters above sea level is equal to its sea level value multiplied by the ratio of atmospheric pressure at the level z to standard sea level pressure"; (*b*) "the precipitable water above the level z is approximately equal to 2.3 $e_z\ 10^{-z/22000}$ cm, where e_z is the vapor pressure in cm at the elevation z in meters"; (*c*) "observations on mountain stations show that dust depletion decreases irregularly from about 4% of the extraterrestrial insolation at 1.5 km to 1% at 4 km." Klein then expresses the transmission coefficient a_z, and the ratio R_z, of diffuse to total radiation with clear sky, both as a function of altitude. One can get the appropriate values from the charts he presents. The total radiation Q received on a horizontal surface is given by:

$$Q = 2.0\ a_z/(1 - R_z) \text{ cal per cm}^2 \text{ per min} \qquad (2.8)$$

One of the frequently occurring problems in ecology, dealing with the solar irradiation of a surface, is the question of how much energy reaches a sloping surface oriented at some angle to the sun. Before dealing with sloping surfaces the geometrical factors affecting the amount of sunlight falling on horizontal surfaces will be discussed. For a horizontal surface outside the atmosphere, the irradiation by direct sunlight can be expressed as

$$Q_o' = \frac{2.0}{r^2} \cos z \qquad \text{cal per cm}^2 \text{ per min} \tag{2.9}$$

where r is the radius vector of the earth which varies with the time of year. Values for r are given in Table 169 of the *Smithsonian Meteorological Tables* (Sixth Revised Edition). The sun's angular zenith distance, z, is given by

$$\cos z = \sin \Phi \sin \delta + \cos \Phi \cos \delta \cos h \tag{2.10}$$

where Φ is the latitude of the station, δ is the sun's declination, and h is the sun's hour angle. The quantities required for computing z from Equation 2.10 are to be found in the *Nautical Almanac.* Values for z may be found directly in Table 170 of the *Smithsonian Meteorological Tables.* By means of Equations 2.9 and 2.10, and the proper integration over time, the values of the extraterrestrial insolation given in Figure 2 were computed. In order to calculate the energy received on a horizontal plane at the earth's surface, one must multiply Equation 2.3 by the cos z factor. This may be written as follows:

$$Q_o = \frac{2.0}{r^2} \exp [\;] \cos z \tag{2.11}$$

where the brackets [ ] contain the factors given in Equation 2.3.

For sloping surfaces it is possible to derive a general relationship for the angle of incidence, i, between the normal to the surface and the ray from the sun, in terms of the following factors: the slope of the surface, β, measured from the horizontal, or in other words, the angle between the normal to the surface and the zenith direction; the zenith angle of the sun, z; the azimuth angle of the sun, α; and the angular deviation in azimuth, η, of the normal to the surface from the true south position. This generalized formula is

$$\cos i = \cos \beta \cos z + \sin \beta \sin z \cos (\alpha - \eta) \tag{2.12}$$

where the solar azimuth angle, α, is given by

$$\sin \alpha = \cos \delta \sin h / \cos z \tag{2.13}$$

Table 170 of the *Smithsonian Meteorological Tables* gives α in graphical form. The geometry of Equation 2.12 applies only to the direct rays from the sun and not to the diffuse radiation from the sky. For south or north facing slopes $\eta = 0°$, for east facing slopes $\eta = -90°$, for west facing slopes $\eta = 90°$. For vertical slopes $\beta = 90°$. The instantaneous irradiation of any slope at the surface would be given by

$$Q = \frac{2.0}{r^2} \exp [\ ] \cos i \tag{2.14}$$

where the brackets again contain the atmospheric transmission factors from Equation 2.3. In order to get the daily total of direct solar radiation on a slope, Equation 2.14 would need to be integrated over time from sunrise to sunset. This can be readily done either graphically or analytically, but the results will not be presented here. The diffuse skylight striking a sloping surface may be important in many instances, particularly for surfaces facing away from the sun and for any surface on a cloudy day. Near noon the diffuse skylight on cloudless days represents about one-sixth of the total radiation falling on a horizontal surface and an increasing fraction of the total toward sunrise or sunset. The best way to get the total irradiation of a slope, direct solar plus scattered solar, is to measure it for the slopes of interest. Values of this type are reported by Orendorff (1954), in connection with architectural design, by Hand (1947), and by Grâfe (1956). In addition to the diffuse scattered light from the sky, there is also the diffusely reflected sunlight and skylight from ground surface which will distinctly contribute to the irradiation of a sloping surface on some occasions. A snow-covered ground will be a situation in which the diffusely reflected sunlight may contribute a significant fraction to a surface above it at almost any orientation. If one compares the total solar radiation, direct plus diffuse sky, plus ground reflected, incident on a vertical north facing surface with direct sunlight incident on clear days, it is found that the north facing wall always receives substantially more radiation due to skylight and reflected light than it would get by direct sunlight. This is particularly true during the winter months at middle latitudes. Grâfe (1956) gives numbers for this ratio for various slopes.

Also of importance in any discussion of solar energy as an environmental factor is the problem of albedo, which by definition is the reflectivity of a surface to visible light. Many albedo measurements have been made and may be found in a number of references including the *Smithsonian Meteorological Tables*, Geiger (1957), and Krinov (1953).

THERMAL RADIATION

The thermal radiation aspects of the physical environment involve certain basic laws of physics which are well-described radiation theory concepts and which are discussed here for purposes of orientation. Some aspects of the thermal radiation environment are conceptually simple while other aspects, such as those dealing with the atmosphere, are somewhat more involved and require more background development for their understanding. The thermal radiation fluxes flowing in the boundary layer adjoining the ground and the atmosphere will be described in detail. By the term "thermal radiation" is meant the emission of electromagnetic waves, heat, light, or radio waves, by virtue of the temperature of an object. Do not confuse the term "thermal radiation" with the term "radiation" or "nuclear radiation" which is commonly found in current scientific literature and which connotes the emission of α particles, β rays, high energy particles and other products of reactions involving the atomic nucleus. The electromagnetic spectrum extends from very short wavelength—high frequency waves in the ultraviolet through the visible—through the infrared or heat region, to the very long

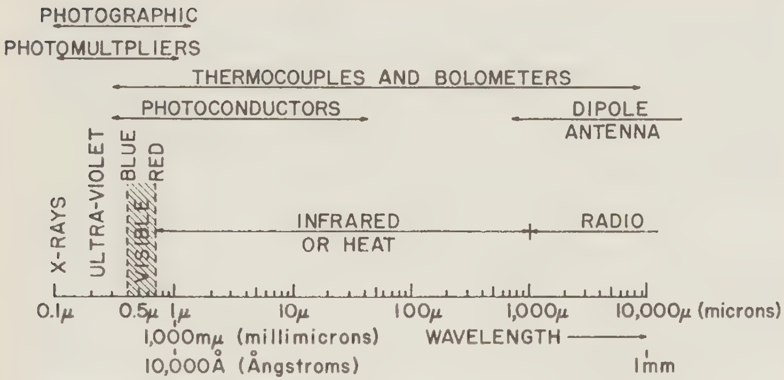


FIGURE 16.—The electromagnetic spectrum and means of detection.

wavelength—low frequency waves in the radio region. As a reminder and a means of orientation, this spectrum is shown in Figure 16. It will be noticed that the visible segment, from 0.3 microns to 0.7 microns (3000 Å to 7000 Å) is a very limited portion of the entire electromagnetic spectrum, limited by virtue of the restricted sensitivity of the human eye to the wavelengths or frequencies we know as violet, indigo, blue, green, yellow, orange, and red. To the high frequency, short wavelength side, there are ultraviolet rays, X rays, and γ rays, all of which are capable of penetrating matter and producing ionization. These short wave radiations may all be detected by means of photographic plates, and this technique will also work throughout the visible and into the infrared as far as 1.2 microns. The infrared or heat radiation segment of the electromagnetic spectrum extends from the visibility limit of the human eye at 0.7 microns (7000 Å) on toward greater wavelengths, until in the vicinity of 1000 microns it is superceded by the radio techniques of the microwave region. The only detectors sensitive to all wavelengths in the infrared are thermocouples and bolometers. Photoconductor detectors such as PbS, PbTe, or PbSe and copper or gold-doped germanium have sensitivities of limited wavelength response. The entire range of usefulness for photoconductors is shown in Figure 16. Most photoconductors must be cooled to liquid nitrogen, hydrogen, or helium temperatures to be highly sensitive. Photoconductors may some day be extended to even longer wavelengths than are now possible. The primary advantage in photoconductors over thermocouples or bolometers is not only their somewhat higher sensitivity, but also their much shorter time constant or response time.

A "warm" object at any temperature above absolute zero will radiate "thermal radiation." This radiation will be characterized by means of its intensity and its wavelength or frequency distribution. The higher the temperature of the radiating object, the greater will be the number of short wavelengths emitted. The lower the temperature of the radiating object, the greater will be the percentage of emitted wavelengths which will fall toward the long wavelength end of the electromagnetic spectrum. This is illustrated for the radiation from a black body in Figure 17. A black body is a perfect radiator or a perfect absorber and can be approximated to a high degree of accuracy by a small opening into

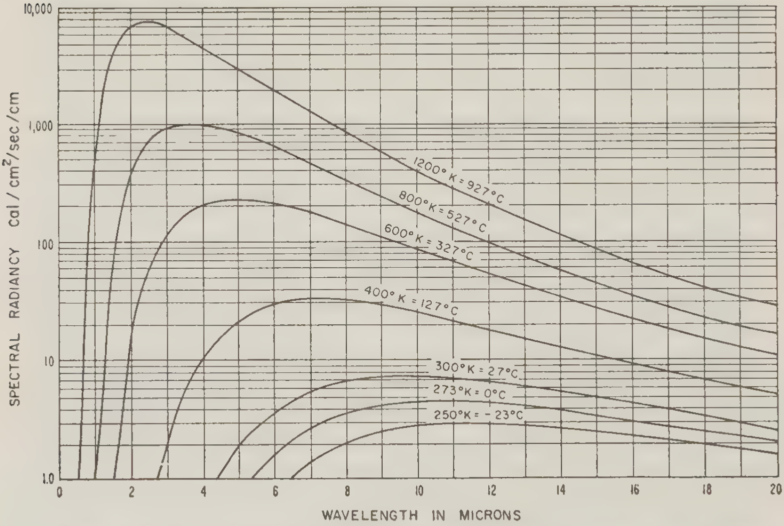


Figure 17.—Spectral radiancy of black body sources at various temperatures as a function of the wavelength.

a cavity enclosure. The behavior of these curves can be recognized from common experience. One often hears it stated that a hot flame is white in appearance, and a cooler flame is more orange or red in color. One has observed the dying embers of a fire change in color from brilliant white, to glowing orange, to dull red, and finally the colors disappear from view but the embers still feel warm to the hand placed nearby. This is simply the process of shifting from one temperature curve at a higher temperature to another curve at a lower temperature, and proportionately toward longer wavelengths having less intensity as the coals cool off. The wavelength behavior of this shift is the well-known Wien displacement law of radiating objects described in terms of the wavelength position, λ_m, of the peak of the radiation curve as follows:

$$\lambda_m T = 2864 \text{ micron degrees Kelvin} \qquad (2.15)$$

where λ_m will be given in microns and T in degrees Kelvin. For an object at a temperature of 286°K or 13°C the radiation emitted will peak in the infrared at 10 microns. If the temperature is doubled, to 573°K, then λ_m will be 5 microns; for $T = 1146$°K, $\lambda_m = 2.5$ microns; for $T = 2292$°K, $\lambda_m = 1.25$ microns; etc. For lower temperatures than 286°K, the radiation emitted will peak at longer wavelengths. The surface of the earth and plants and animals radiating with surface temperatures in the neighborhood of 290°K will have a wavelength peak in the vicinity of 10 microns in the infrared. Thus all objects, animate or inanimate, occurring in nature on the planet earth at these temperatures, will exchange infrared radiation with their surroundings.

The total intensity, J_{BB}, of black body radiation radiated will be equal to the area under the spectral curve. Another well-known law, namely the Stefan-Boltzmann radiation law, describes the temperature behavior of this total energy to be

$$J_{BB} = \sigma T^4 \qquad (2.16)$$

where $\sigma = 5.67 \times 10^{-5}$ erg per cm^2 per °K^4 per sec or 1.32×10^{-12} cal per cm^2 per °K^4 per sec, known as the Stefan-Boltzmann constant. At a temperature of 20°C or 293°K one calculates $J_{BB} = 9.73 \times 10^{-3}$ cal per cm² per sec or 0.584 cal per cm² per min. If one were to double the absolute temperature to 586°K the total energy radiated would increase sixteenfold. If an object radiates not as a black body but as a so-called "gray body," then, providing the emissivity, ϵ, is constant with wavelength, the total intensity of the radiated radiation will be given by the following relationship:

$$J_{GB} = \epsilon\sigma T^4 \tag{2.17}$$

The shape and strength of the curves given in Figure 17 are described by means of the Planck radiation law which may be written:

$$J_\lambda = C_1\lambda^{-5}(e^{C_2/\lambda T} - 1)^{-1} \tag{2.18}$$

where $C_1 = 3.703 \times 10^{-5}$ erg cm² sec^{-1}, $C_2 = 14{,}330$ micron°K, λ is wavelength in microns, and T is temperature in °K. The other radiation laws may be derived from this one.

Another law of radiation states that a good absorber is a good emitter at the same wavelength and temperature. A black object which absorbs efficiently certain wavelengths incident upon it must also emit efficiently at the same wavelength, for if it did not, it would heat up indefinitely. A mirror surface will absorb very little radiation and in turn will radiate very poorly. The surface of the earth is mostly black in the infrared and therefore absorbs and radiates essentially as a black body. The leaves of plants have low reflectivity in the infrared, about 5% as shown by Gates and Tantraporn (1952), and therefore are 95% absorbers and in turn radiate 95% of black body radiation. Snow is "black" at infrared wavelengths and absorbs nearly all infrared radiation incident upon it and radiates as a black body surface would. Therefore, for solid objects such as these, where the emissivity is known, it is a simple matter to compute, by means of Equation 2.16, the amount of energy radiated if one knows the surface temperature. If one is in a room with walls at 20°C the radiant heat load will be 0.6 cal per cm² per min as computed from Equation 2.16. This is shown in Figure 10 as a horizontal line.

The atmosphere, containing various gases, is neither "black" nor entirely transparent at infrared wavelengths. The character of the infrared absorption spectrum of the atmosphere is shown in both Figures 3 and 6. There are some absorption bands which are "black" or opaque for the depth of the entire atmosphere. The intense absorptions due to water vapor in the atmosphere are located at 1.4, 1.9, 2.4, 2.7, and 6.3 microns, and from 22 microns continuously on toward longer wavelengths. There are weaker bands due to water vapor at 0.7, 0.77, 0.82, 0.93, and 1.1 microns. There are also strong absorptions due to carbon dioxide located at 4.3 and 15.0 microns and a weaker one at 2.7 microns overlapping with the water vapor absorption at this wavelength. The only other really significant absorption in the infrared is the band at 9.6 microns due to ozone. The spectrum also contains very weak absorptions not shown in Figure 6, which are produced by carbon monoxide, nitrous oxide, methane, and heavy water. The atmosphere absorbs radiation of these wavelengths from the sun and from the ground and will, in turn, reradiate energy at these wavelengths in strengths proportional to the absorption bands and the temperature of the atmosphere.

The actual process of absorption and reradiation by the atmospheric gases is very complicated, since the process is distributed throughout the atmosphere from top to bottom over vast pressure and temperature differences. The meteorologist is concerned with how various levels of the atmosphere are heated or cooled by the radiation process. For purposes of biological environments, one is primarily concerned with the strength of the radiation fluxes at the earth's surface. The nature of the downward stream of radiation coming from the atmosphere toward the surface and the upward stream from the ground can be appreciated from Figure 6. If the ground surface temperature is 15°C or 288°K, the upward stream of radiation will be simply that calculated from Equation 2.18 and will have a continuous envelope with wavelength as marked in the middle part of the figure. In order to calculate the downward stream from the atmosphere, one must consider the selective radiation by the atmospheric gases. The radiation emitted by the water vapor and carbon dioxide largely comes from the troposphere—that portion of the atmosphere between the ground and the stratosphere. If the troposphere were indeed opaque, it would radiate essentially as a black body at a temperature of about 263°K and would exhibit the radiation curve in Figure 6 marked "black body emission 263°K." However, since the atmosphere is distinctly "gray" and absorbs and radiates only in certain discrete bands, it is necessary to apply this band emissivity to the continuum curve to get the crosshatched areas shown under the curve. It will be noticed that these crosshatched emission zones appear almost as mirror images of the absorption bands located at the top of Figure 6. The one absorption band in the spectrum that cannot be considered as tropospheric absorption or emission is that at 9.6 microns due to ozone. This is because most of the ozone in the atmosphere is located in the stratosphere at a height of 20 to 30 kilometers and at a temperature of approximately 235°K. In order to estimate the strength of the radiation flux originating with the ozone in the stratosphere, the magnitude of the absorption band at 9.6 microns is applied to the black body emission from an object of 235°K to give the emission peak located at 9.6 microns in the center diagram. So it will be noticed that the downward stream of infrared thermal radiation from the atmosphere is quite variable with intensity with wavelength but that a significant proportion of the radiant energy is distributed among wavelengths greater than 13 microns. The 15 micron carbon dioxide band, and the so-called "pure rotation" band of water vapor extending throughout this long wavelength region and on out to microwave wavelengths, are both extremely important contributors bathing the earth's surface with a warm flux of radiation. A portion of this spectrum has been observed by Sloan, Shaw, and Williams (1955), with a low-resolution infrared spectrometer pointed toward the night sky. This spectrum is displayed in the lower portion of Figure 6.

The spectral properties of the thermal emission from the sky or atmosphere are important for two reasons: the first being the dependence of the downward thermal radiation stream on the atmospheric composition, concentration, and temperature; and the second being the fact that plants and animals may have different responses to the various spectral regions. Not a great deal is known concerning the response of various organisms to infrared radiation, although it is entirely likely that most organisms absorb infrared radiation uniformly and fairly completely, thereby converting the radiation to heat within the surface cells.

It is possible to express the magnitude of the downward flux of thermal radiation from the atmosphere as a purely analytical expression. The mathematical derivation of this analytical expression is very involved and will not be given here. A good description of the derivation, as well as its presentation in terms of the Elsässer radiation chart, may be found in Haltiner and Martin (1957) and in Berry, Bollay, and Beers (1945). Elsässer (1960) has recently revised the radiation chart in order to make it more accurate. The radiation chart is a convenient graphical method for obtaining a solution to the analytical expression for the strength of the radiation flux in terms of the moisture and temperature profile with height in the atmosphere. However, its use depends upon information from meteorological soundings which give the desired profiles. In addition the percentage of cloud cover, the cloud type, and the temperature of the base must be known. With all of this information available, a reasonably accurate computation may be made of the downward stream of thermal radiation from the atmosphere at any given time and place. Another form of radiation chart is given by Robinson (1950) and also one by Yamamoto (1952) and Yamamoto and Sasamori (1954), the latter claiming somewhat greater accuracy and pointing out certain corrections necessary in the Robinson chart.

Before analytical expressions which lead to the use of radiation charts had been derived, Ångstrom (1915) and Brunt (1932) had arrived at empirical expressions for the downward stream of thermal radiation from the clear sky. The Ångstrom formula is

$$R_o = \sigma T^4(0.81 - 0.2411^{-0.052e}) \tag{2.19}$$

where σ is the Stefan-Boltzmann constant, T is the surface air temperature, and e is the partial pressure of water vapor at the surface expressed in millibars. The Brunt formula is

$$R_o = \sigma T^4(0.44 + 0.08 \sqrt{e}) \tag{2.20}$$

These empirical formulas obviously do not agree for the condition of $e = 0$, i.e., a dry atmosphere, when the radiation would be primarily determined by the carbon dioxide content of the atmosphere. Both formulas give somewhat high results for absolutely dry atmospheres, since the true amount of radiation due to carbon dioxide in the normal concentration of 0.03% is 0.185 σT^4. Also it is apparent that formulas such as these do not in any way take account of the moisture and temperature profile in the atmosphere and the time variation of these variables at any level other than the surface. The formulas work as well as they do because much of the moisture in the atmosphere is tropospheric and is concentrated near the surface. However, it is often the situation that only the surface conditions are known or available, and for this reason these equations are very useful in making rough estimates of the downward thermal radiation flux from the clear sky.

In order to take into account the influence of clouds, Ångstrom and Asklof (1920) have given the radiation from overcast sky expressed in terms of the clear sky radiation, R_o, of Equation 2.19 as follows:

$$R = \sigma T^4 - \lambda(\sigma T^4 - R_o) \tag{2.21}$$

where λ is a parameter which depends upon the height of the cloud base, being 0.17 at 2.0 km, 0.38 at 5.0 km, and 0.45 at 8.0 km. It is obvious that further refinement of this formula is required to take into account partially clouded skies.

Using a value for σ of 1.32×10^{-12} cal per cm^2 per °K^4 per sec or 0.793×10^{-10} cal per cm^2 per °K^4 per min, one calculates the values given in Table III, using Equation 2.19.

For orientation purposes it is valuable to consider the typical values of vapor pressure for various temperatures and relative humidities shown in Table IV.

The exponential behavior of the radiation with vapor

TABLE III

VALUES OF R_o (IN CAL PER CM^2 PER MIN)

	250°K	300°K	320°K
0 mb	0.176	0.363	0.469
10 mb	0.231	0.477	0.617
40 mb	0.252	0.522	0.700

TABLE IV

VAPOR PRESSURE VALUES

(IN MILLIBARS)

Relative Humidity	T					
	0°C	10°C	20°C	30°C	40°C	50°C
100%	6.1	12.3	23.4	42.4	73.8	123.4
50%	3.1	6.2	11.7	21.2	36.9	61.7

pressure in Equation 2.19 shows up in the calculations wherein the changes in radiation flux are greatest for small values of vapor pressure or dry atmosphere. This is because the water vapor absorption or emission bands rapidly saturate as the amount of water vapor is built up.

RADIATION FLUXES WITHIN SPECIFIC ENVIRONMENTS

It is of interest to compute the energy fluxes flowing within certain hypothetical environments which are nevertheless based on reality. By means of the formulas given above, the values shown in Table V have been worked out for the conditions listed. In Table V, T_A represents the air temperature near the surface, and T_G the surface temperature of the ground. The total direct sunlight plus scattered skylight on a horizontal surface was taken to be 1.0 cal per cm^2 per min, which is a typical clear sky, summer day value for middle latitudes. Larger values, which do occur, will increase the heat load above the values given here. Values of energy flux or heat flow are not of significance by themselves but in relationship to something else. For the meteorologist they

TABLE V

RADIATION REGIMES FOR CERTAIN HYPOTHETICAL ENVIRONMENTS

Conditions	T_A (°C)	T_G (°C)	Albedo (%)	Vapor Pressure (mb)	Solar Sky	Refl. Solar	Downward Long Wave	Upward Long Wave	Net Flux	Total Absolute Flux	Absorption by Horiz. Leaf
					(cal per cm² per min)						
Sand dunes											
Clear day	30	40	30	36.1	1.00	−0.30	0.556	−0.792	0.464	2.648	2.400
Cloudy day	30	30	30	36.1	0.57	−0.17	0.666	−0.696	0.370	2.102	1.946
Clear night	15	10	—	15.2	—	—	0.434	−0.528	−0.094	0.962	0.932
Cloudy night	15	15	—	15.2	—	—	0.534	−0.557	−0.023	1.091	1.058
Open clearing in forest with grass											
Clear day	30	40	7	36.1	1.00	−0.07	0.556	−0.792	0.694	2.418	2.210
Bog with lake											
Clear day											
Over water	30	25	9	36.1	1.0	−0.09	0.556	−0.643	0.823	2.289	2.089
Over mat	30	40	2	36.1	1.0	−0.02	0.556	−0.792	0.744	2.368	2.167
Interior of deciduous forest											
	T crowns										
Clear day	23	20	—	—	0.05	0	0.637	−0.607	0.080	1.294	1.247
Clear day	35	30	—	—	0.05	0	0.738	−0.696	0.042	1.484	1.431

are of importance for estimating the temperature changes which a parcel of air may undergo, or for forecasting the occurrence of frost. For the biologist the interest may center on the organisms themselves and the effects of the heat load on the organism. As an example of the usefulness of the calculations being made here, the last column in Table V gives the total radiant energy absorbed by the upper and the lower surfaces of a horizontal leaf suspended in the environments given in the table. The heat load for the leaf oriented in other directions could also be calculated, however the horizontal position is the most convenient. Some of the values of the heat load given in Table V are also reproduced in Figure 10. It should be noticed that the total heat load on two surfaces of the horizontal leaf is always considerably greater than the direct sunlight would be on an upward facing horizontal surface during the daytime. The maximum heat load encountered will be approximately 2.40 cal per 2 cm^2 per min. If all of this energy were to go into increasing the temperature of the leaf, the temperature of the leaf would reach a fantastically high value. This temperature increase can be calculated by means of the following relationship:

$$\frac{\Delta T}{t} = \frac{Q}{t}\frac{A}{Mc} \qquad (2.22)$$

where Q/t is the radiation flux in cal per cm^2 per min, M/A is the mass of the leaf per unit area and is taken to be 0.02 grams per cm^2, c is the specific heat taken to be 0.88, and $\Delta T/t$ is the time rate of temperature change. If $Q/t = 2.40$ cal per 2 cm^2 per min, then $\Delta T/t = 136$°C per min. If $Q/t = 2.00$, then $\Delta T/t = 114$°C per min. Obviously these are ridiculous answers. Several events occur which have not been taken into account. The leaf reradiates most of this energy, some of the energy is convected away, and transpiration works to keep the temperature down. For the present the convection loss will be ignored, but it will be dealt with in a later chapter. It is easy to calculate the temperature the leaf would reach if it lost all of the heat load by reradiation. Assume the leaf to radiate 95% as effectively as a black body.

Then for equilibrium between heat load and reradiation one must have:

$$0.95\ \sigma T^4 = 0.753 T^4 = \frac{Q}{t} \tag{2.23}$$

Using $Q/t = 2.40$ cal per 2 cm² per min, one gets $T = 352$°K or 79°C which again is too high a temperature. This obviously means that the other mechanisms for carrying away the energy are playing a significant role. Ehlers (1915) has shown that leaf temperatures may run 2° to 10°C above air temperatures during winter conditions. For the first case given in Table III, which has an air temperature of 30°C and is the situation giving rise to a heat load of 2.40 cal per 2 cm² per min, one can assume the leaf temperature to not exceed 40°C and compute the transpiration rate required to maintain this temperature. The leaf will radiate at this temperature 0.791 cal per cm² per min from one surface and 1.582 cal per 2 cm² per min from two surfaces, and the remaining 0.818 cal per 2 cm² per min must be given up through transpiration and convection. If the latent heat of evaporation for water at an initial temperature of 40°C is taken as 600 cal per gram, then the leaf will need to transpire $0.818/600 = 0.00136$ grams of water per 2 cm² of surface area per min or 0.0818 grams per 2 cm² per hr. If this value is unreasonably high for the transpiration rate, then it means that the leaf temperature must rise to a higher maximum value or a considerable amount of heat must be carried away by convection. If the leaf temperature were to rise to 50°C, then it will radiate 1.720 cal per 2 cm² per min from two surfaces. The residue of heat would amount to $(2.400 - 1.720) = 0.680$ cal per 2 cm² per min, and the transpiration rate would be 0.00113 grams of water per 2 cm² per min or 0.068 grams per 2 cm² per hr. The leaf temperature will need to rise considerably higher before the transpiration rate is reduced to a small value, or considerable convection must occur. As shown in Table V and Figure 10, various other lesser heat loads may frequently occur, and the transpiration rate in these cases will be correspondingly less. However, it should be emphasized that for an estimate of transpiration rates, the complete heat load on a leaf must be arrived at either through calculation or measurement. To use only the incident solar radiation component is to obtain only part of the answer and will give a gross underestimate of the effects. Convection effects will be considered in a later chapter.

Chapter 3

Radiation Instruments

INTRODUCTION

Instruments for the detection and measurement of solar and thermal radiation may be classified and described according to their wavelength range of sensitivity and their mode of operation. Not included in the present discussion are spectrometers, interferometers, or other instruments with dispersive characteristics. The instruments to be described here either receive and detect all wavelengths of energy simultaneously, or broad wavelength bands of energy as passed by a filter. Table VI gives a general classification of radiation instruments. The classification is according to the wavelength response of the instrument. Any instrument which has a glass dome or cover is limited in its wavelength response to the range 2800 Å to 30,000 Å. Crystal quartz will extend the range 2000 Å to 40,000 Å and fused quartz to a somewhat shorter wavelength on the ultraviolet end. When very thin polyethylene is used as a dome, it essentially transmits all wavelengths of energy, both the short wave and the long wave, and therefore is not classified as an instrument with limited wavelength response.

The instruments which measure the short-wave sunlight and skylight energy are referred to generally as pyrheliometers, pyranometers, pyrgeometers, actinometers, and solarimeters, etc. An example of this type of instrument is shown in Figure 18 (see opposite p. 78). If the instrument is self-recording, the term "graph" will replace the term "meter." Instruments which measure the total flux of energy of all wavelengths received on a single surface will be termed radiometers. If the instrument has two receiving surfaces, an

TABLE VI
Radiation Instruments

A. Direct Sunlight and Skylight Response (2800–30,000 Å)

1. Eppley pyrheliometer
2. Moll-Gorczynski solarimeter
3. Stern pyranometer
4. Robitzsch bimetallic actinograph
5. Bellani pyranometer

B. Total Radiation Response

a. Ventilated net radiometers or balance meters

1. Franssila-Suomi net radiometer or balance meter
2. Gier and Dunkle net radiometer or balance meter
3. Kew Observatory net radiometer or balance meter
4. Courvoisier net radiometer or balance meter

b. Nonventilated exposed surface instruments

1. Hofman heat-compensated net radiometers or balance meters
2. Ångstrom compensation pyrheliometer and pyrgeometer
3. Silver disc pyrheliometer
4. Monteith net radiometer or balance meter
5. Stoll-Hardy radiometer
6. Frigorimeter

c. Nonventilated enclosed surface radiometers

1. Schulze radiometer
2. Suomi "economical" net radiometer or balance meter
3. Funk net radiometer or balance meter
4. Fritschen and van Wijk net radiometer or balance meter

upward facing and a downward facing, and responds to the difference signal between the two surfaces, then it may be termed a net radiometer or balance meter. An example of a net radiometer or balance meter is shown in Figure 20 (see opposite p. 78).

DIRECT SUNLIGHT AND SKYLIGHT RESPONSE

Eppley pyrheliometer

The Eppley pyrheliometer is the most widely used solarimeter in the Western Hemisphere, having been the standard U.S. Weather Bureau instrument for recording solar radiation. This instrument consists of a thermopile constructed from an alloy, 60% gold and 40% palladium, against another alloy, 90% platinum and 10% rhodium. The "hot" junctions are in thermal contact with a blackened silver ring which is surrounded by a concentric ring painted white with magnesium oxide with which the "cold" junctions are in thermal contact. Usually 16 or 50 junctions are employed, depending upon the desired sensitivity. The receiving assembly is enclosed in a spherical glass bulb which is hermitically sealed and which limits the wavelength response to wavelengths shorter than 3.5 microns. The voltage response of the instrument is nearly linear with incident radiant flux and amounts to about 8 millivolts per cal per cm^2 per min for the 50 junction model. The time constant is of the order of 20 seconds and the resistance 100 ohms. The instrument is manufactured by the Eppley Corporation, Newport, Rhode Island. For a description of the instrument see Kimball and Hobbs (1923).

The instrument, when used fully exposed to the upward hemisphere, will record both the direct sunlight and the scattered skylight. If the instrument is inverted to the downward facing position it can be used to record reflected sunlight and skylight from the ground surface. If the instrument is to be used for measuring only the scattered skylight, it will be used in the upward facing position with the direct sun obscured by means of a metal disc or band mounted on an arm so that its shadow will always fall across the receiving element. The Eppley pyrheliometer can be operated for long periods of time in the field with very little attention required. The signal from the instrument can be recorded with any standard pen and ink electrical potentiometer.

Moll-Gorczynski solarimeter

The basic principle of the design of this instrument is similar to the Eppley, however it is different in detail and of independent invention. The receiver is a blackened, horizontal, rectangular receiving area backed with 14 manganin-constantan thermojunctions in good thermal contact. The receiving area is surrounded by a heavy brass block with which the "cold" junctions are in thermal contact. This unit is covered with double concentric hemispherical glass domes of 2 mm thickness and of diameters 30 and 50 mm. The purpose of the double glass domes in this instrument is to reduce the transfer of heat from the receiving surface to or from the outer dome by convection. The glass domes are cemented into circular grooves in the brass casting and when subjected to severe weather conditions sometimes loosen and leak. They can be resealed by use of "Bostik glazing compound," and this has proven more weather resistant than the original cement used by the manufacturer. The sensitivity is about 8 millivolts per cal per cm^2 per min, the resistance is 10 ohms, and the time constant is about 10 seconds. The instrument is considered somewhat more accurate than the Eppley instrument and, except for the difficulty with the domes, may be more rugged. Naturally the instrument is more commonly used in Europe where it is manufactured by Kipp and Zohnen in Delft, Holland.

Stern pyranometer

The Stern pyranometer is shown in Figure 18 (see opposite p. 78). As illustrated here it has 32 thermocouple junctions which make contact with alternated black and white receiving surfaces. The black absorb the sunlight and the white reflect the sunlight. The output of the instrument is between 1.5 and 2.0 millivolts per cal per cm^2 per min. The instrument has an internal resistance of about 5 ohms. A single glass dome shields the receiving surface from strong disturbances caused by wind. The instrument therefore only records the visible and near infrared wavelengths in the range transmitted by glass. The Stern pyranometer can be obtained from Philipp Schenk, 40 Voltagasse, Vienna 21, Austria.

Robitzsch bimetallic actinograph

This instrument receives radiation on a long thin blackened metal strip about 8.5 mm by 1.5 mm. Two long thin strips, paralleled to the black strip and to either side, are painted white in order to reflect the radiation. The three strips are connected so as to cause the mechanical distortion of a lever which actuates a pen arm. The pen rests on a paper chart which is driven by a spring. The instrument is designed in such a way that ambient temperature changes are compensated and produce no deflection of the pen. The instrument will register the daily totals of solar radiation and sunlight to an accuracy of 5% to 10%. Instantaneous values are somewhat less accurate, especially since the instrument has a long time constant, of the order of 2 minutes, and will not respond to quickly changing amounts of sunlight.

The Robitzsch bimetallic actinograph was first produced by Fuess in Berlin-Steglitz who still continue its manufacture. Other models are now made by the Italian firm Societa Italiana Apparecchi di Precisione in Bologna, by Casella in London, and by Belfort Meteorological Observatory in Baltimore. The Belfort instrument employs only two long thin metal strips rather than three. References describing the use and calibration of the instrument include Robitzsch (1932), Morikofer (1939), Courvoisier (1946), Schüepp (1952), and Blackwell (1953), among others. The instrument has had a great deal of appeal since it can be used in the field remote from power supplies and essentially left without attention for several days.

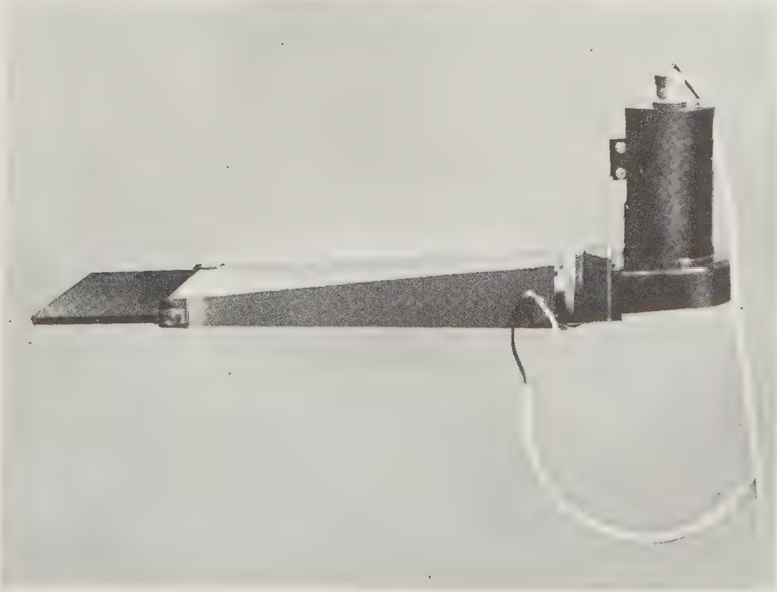
Bellani pyranometer

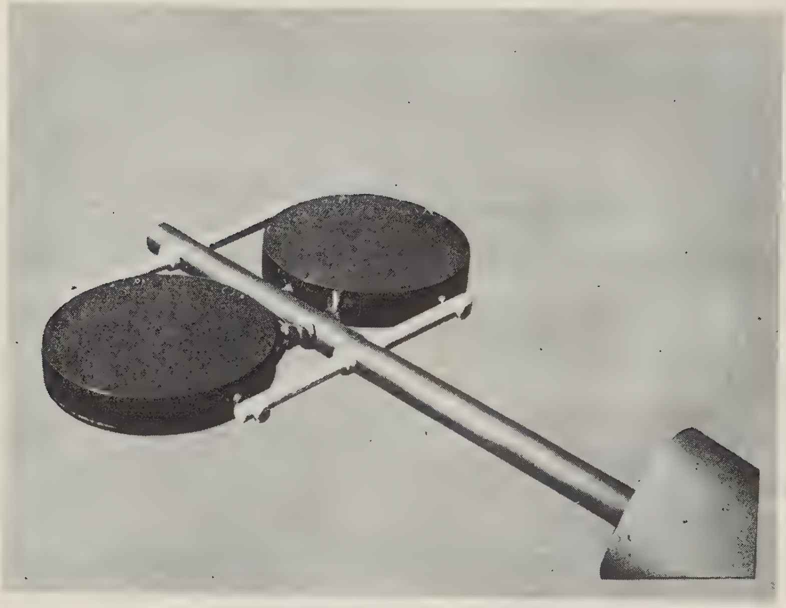
This is a simple self-registering radiation instrument for giving the daily totals of sunlight and skylight of wavelengths

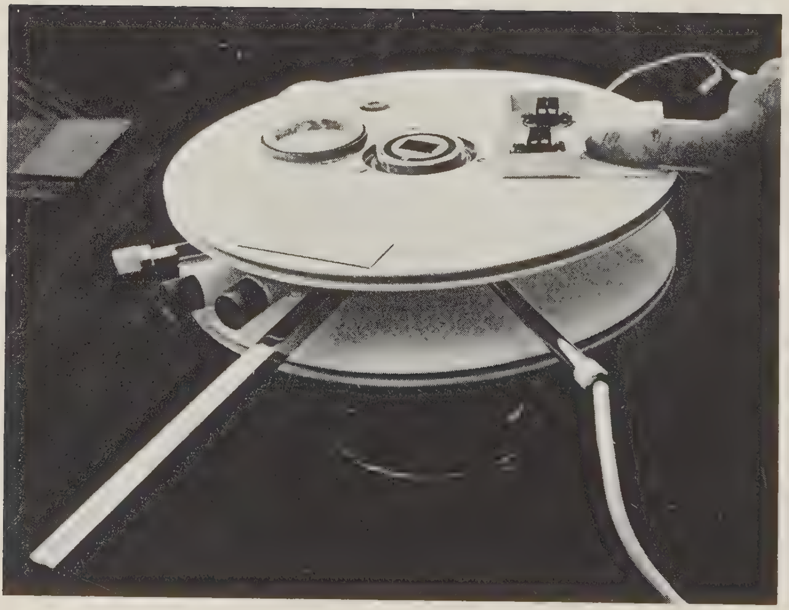
FIGURE 18 (*top*).—Stern pyranometer for observing the direct solar plus sky radiation and meter for direct reading of the output in cal per cm^2 per min. The glass dome limits the response to radiation wavelengths shorter than 3 microns.

FIGURE 20 (*bottom*).—Franssila-Suomi net radiometer or balance meter for measuring the difference between the downward and the upward streams of radiation. All wavelengths are detected.

cal/cm²/min
00
02
04
06
08
12
14
16
0
20
40
60
80
100







shorter than 3.0 microns. The instrument is particularly useful for field work involving biological climatology studies. A cross section of the original spherical Bellani pyranometer is shown in Figure 19. It is comprised of two concentric glass spheres, *G*, mounted on the end of a calibrated glass tube, *A*. The reservoir within the inner sphere which is filled with alcohol is open to the calibrated tube through a smaller channel, *C*. The inner glass sphere is completely encased in a metal surface, *R*, which is coated with a black paint and acts as the receiver surface. The receiver surface becomes warm and transfers heat to the volatile alcohol which in turn vaporizes, filling the tubes *C* and *T* with vapor which then condenses in the bottom of *C*. The amount of liquid alcohol accumulated in the tube, *T*, which is calibrated, is a measure of the total short wave radiation incident on the spherical receiving surface. Because the receiving surface is spherical, it receives and registers the radiation received from the sun and sky as well as the sunlight reflected from the ground. For this reason it does not give the radiation received on a horizontal surface. In order to eliminate convective and conductive loss of energy to the outside, the space between the two glass spheres is evacuated.

A new design of the instrument by Courvoisier and Wierzejewski (1954) employs a black horizontal surface as receiving element mounted inside a cylindrical glass envelope. The black receiver is shielded on the sides by a metal cylinder. This instrument only responds to the total sunlight and skylight received on the horizontal surface and does not include reflected radiation. Except for the cylindrical geometry rather than spherical, the principle of operation is precisely the same

---

FIGURE 21 (*top*).—Hofman heat-compensated net radiometer or balance meter. Two instruments are shown since each disc is a complete instrument. All wavelengths are detected.

FIGURE 23 (*bottom*).—Schulze radiometer with the polyethylene dome removed, exposing the blackened receiver. The small box at the right shows two pairs of thermopiles. The downward facing half of the instrument is identical with the upward facing half.

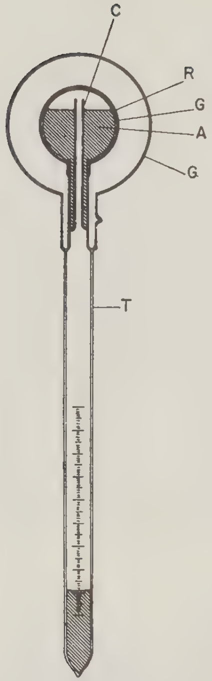


FIGURE 19.—Bellani pyranometer. The globe, *G*, is filled with alcohol, *A*, which vaporizes, passes through the capillary, *C*, and condenses in the tube, *T*. The amount of alcohol accumulated in the tube is a direct measure of the amount of radiation received on the blackened spherical receiving surface, *R*. The glass dome limits the response to radiation wavelengths shorter than 3 microns.

as with the earlier version. The instrument is readily reset by simply inverting it, thereby permitting the alcohol to refill the reservoir at the top. Some residual alcohol will remain in the tube, *T*, but the level is simply noted and the difference in readings during the period of measurement is recorded. The Bellani pyranometers give the daily totals of sunlight and the skylight within a 3% accuracy. The old model and the new model are available from the Physikalisch-Meteorologischen Observatorium, Davos, Switzerland.

TOTAL RADIATION RESPONSE

Black receivers, which are not enclosed within a glass cover, are capable of responding to all wavelengths of radiation, including the long-wave infrared. However, an exposed black surface is subject to heat transfer to the air by convection and conduction and to the effects of moisture on the surface. In order to reduce or eliminate these disturbing forces several different arrangements are employed. The receiving surface may be ventilated at a constant rate, or the sensitive elements may be made self-compensating to wind effects by being shielded from the wind by means of a narrow aperture, or shielded with a dome of material other than glass which may be transparent to infrared as well as to visible wavelengths. Essentially two types of instruments exist: effective pyranometers, which measure the total incident flux of radiation, and radiation balance meters, which measure the difference between two streams of radiation.

Ventilated net radiometers or balance meters

These instruments operate with the horizontal receiving element mounted in a stream of air forced across the surface by means of a blower. A typical instrument is shown in Figure 20 (see opposite p. 78). The blower motor should be mounted as far from the sensitive element as possible. The air is directed to the receiving surface from the blower through a vent, nozzle, or wind tunnel. The forced ventilation should be strong in order to prevent external wind disturbances. The heat sensitive element is simply a thermopile which is wound on a support of insulating material and then is coated over with black paint or lacquer. The paint now recommended for all radiation instruments is Parson's Optical Black Lacquer from Parson's Ltd., London. This lacquer has an emissivity of 98% or greater over the visible and infrared regions of the spectrum to wavelengths well beyond 10 microns. The thermopile is often made by winding a constantan wire about the plastic holder until a sufficient number of turns is reached. Then half the plate can be dipped into a copper sulphate solution and copper plated onto the lower half of each turn. This then forms a series of copper constantan junctions on the two sides along the center line. The thermopile can then be coated with black and is ready to mount. Several versions of this instrument have been designed and built in various parts of the world. Some of these designs will now be described.

Franssila-Suomi net radiometer or balance meter

This instrument, shown in Figure 20, is described in detail by Suomi, Franssila, and Islitzer (1954). The instrument is well designed in that the blower housing is small and is located as far as possible from the receiving surface. The ventilation velocity is high. Two additional features are unique with this instrument. The vent contains a vane which can be adjusted in order to equalize the flow of air over the upper and lower surfaces. In addition there is an electric heater element on either side of the sensitive element which can be used to furnish a sensitive control over the ventilation or convective transfer of energy. The instrument constant is 0.5 to 0.6 cal per cm^2 per min per millivolt and a recorder with 4 to 5 millivolts full scale is required.

Gier and Dunkle net radiometer or balance meter

This instrument was first described by Gier and Dunkle (1951). The thermopile consists of 300 silver-constantan junctions wound on a bakelite plate of area 115 mm by 115 mm. The thermopile has bakelite plates above and below it, each of which is covered with aluminum foil. This heat flux unit is then mounted on the end of a short vent and is much too close to the blower housing unit for completely satisfactory operation. When used as a total hemispherical radiometer, one surface of the heat flux plate is blackened and the other surface is left as shiny aluminum. For this use, an aluminum shield is mounted outside of the shiny surface with sufficient space in between for the forced ventilation to pass. When used as a balance meter or net exchange radiometer, both surfaces of the heat flux plate are blackened and no radiation shield is used externally. When employed in this way, the thermopile measures the difference between the heating of the two surfaces. The general sensitivity of the instrument is about 5 mV per cal per cm^2 per min and the response time is about 10 seconds. The Gier and Dunkle instruments are manufactured by Beckman and Whitley, San Carlos, California.

Kew Observatory net radiometer or balance meter

This instrument is very similar to the other models described above. It has gone through a series of developmental stages during the last few years. For several years Kew Observatory has constructed their own thermopiles, but more recently they are purchasing the two-inch diameter heat flow unit from Joyce, Loebl and Co., Ltd., Newcastle. The heat flow unit is mounted on the end of a long slender ventilating throat and the blower motor is capable of producing approximately 20 m per sec flow of air. This flow rate is necessary in order to prevent external winds from disturbing the reading. The sensitivity of the instrument is approximately 55 mV per cal per cm^2 per min from a thermopile comprised of nearly 900 thermojunctions.

Courvoisier net radiometer or balance meter

The Courvoisier balance meter is a much more elegantly constructed instrument than any of the other types, but it is also more expensive. It is better designed from the standpoint of the wind tunnel than the other instruments of this type. This instrument is manufactured by the Physikalisch-Meteorologischen Observatorium, Davos, Switzerland. It is briefly described in a paper by Courvoisier (1950).

Nonventilated exposed surface instruments

These instruments have the receiving element exposed directly to the stream of radiation without a protective cover and without a constant stream of ventilation. They may or may not be compensated for the influence of convection on the receiving element. Compensation may be accomplished by having dual receiving elements and differences in response are balanced out by heat input on one or both of the elements. Certain of these instruments, such as the Monteith net radiometer or balance meter, operate on the principle that the two receiving elements are each subjected to the same degree of convective cooling to a first approximation. The Stoll-Hardy radiometer has a single receiving element which is disturbed seriously by wind and so is usually used in still air.

Hofman heat-compensated net radiometers or balance meters

The instrument designed and described by Hofman (1952) is shown in Figure 21 (see opposite p. 79). Two instruments are actually shown, each disc being a single instrument. A very similar radiometer has been used by Wagner in Hamburg and also by the Dutch in de Bilt. A heating coil is imbedded just beneath the black surface on both the upper and the lower sides, and thermopiles or thermistor elements make thermal contact with the inner side of the two surfaces. The heat input is adjusted until the temperature of the two surfaces is equal. The assumption is made that the convective heat losses will be equal from the upper and lower surfaces under this condition. This assumption is not correct and considerable error can result, although proper calibration can reduce this error. Another difficulty regarding the instrument is the power requirement and the problem of regulating the heat flow. However the Hofman radiometer or balance meter has been successfully used by Baumgartner (1956a) for studies of the radiation balance in and above forests in Bavaria. Somewhat greater accuracy and convenience can be achieved by using the radiometer in pairs. When this is done, one need not maintain the temperatures of the surfaces at equal temperatures but need only to put in a fixed and measurable quantity of heat, H, to the upper surface of one radiometer and to the lower surface of the other. Let R be the net downward or upward stream of radiation to be measured. The thermopile will indicate a voltage difference between the surfaces for the radiometer on the left of V_l and for the radiometer on the right of V_r where V_H, V_R, and H are all measurable quantities and hence R can be determined. In this form the

$$\begin{aligned}
V_l &\propto (R + H) \\
V_r &\propto (R - H) \\
V_R &= V_l + V_r \propto 2R \\
V_H &= V_l - V_r \propto 2H \,, \qquad (3.1) \\
\frac{V_H}{V_R} &= \frac{H}{R} \qquad R = \frac{V_R}{V_H} H
\end{aligned}$$

balance meter has been used for work in forest meteorology by Baumgartner (1956b) and also has been used for a radiation radiosonde in Germany by Pohl (1955). The instrument measures all wavelengths, visible and infrared, of the incident radiation.

Ångstrom compensation pyrheliometer and pyrgeometer

This instrument was first designed and reported in the literature by Ångstrom (1905). In the pyrgeometer the sensitive element consists of four manganin strips. Two of these strips are blackened and the other two are polished and thereby reflecting. The polished strips remain at ambient air temperature; the blackened strips are exposed to the night sky and will cool by radiation exchange with the sky. Their temperature is maintained at ambient by passing a heating current through the strips. Knowing the temperature of the strips and their emissivity, one can calculate the strength of infrared radiation reaching the instrument from the sky. The pyrgeometer can only be used at night and is subject to disturbances from the wind.

The Ångstrom compensation pyrheliometer can be used in the daytime and employs two blackened manganin strips which are exposed to the sun through an extended tube. One strip is shielded from the sun with a shutter and carries an electric current of sufficient strength to heat it to the same temperature as the strip receiving the incident sunlight. Then the situation is reversed by turning the shutter, covering the other element, and a measurement made under this condition. These two instruments measure all wavelengths of energy since the sensitive elements are freely exposed to the incident radiation.

Silver disc pyrheliometer

This instrument was developed by Abbot of the Smithsonian Institution and has been used as a secondary standard for solar radiation measurements for many years. A silver disc is located at the lower extremity of an extended tube containing a series of apertures or diaphrams. A very accurate thermometer is inserted into the silver disc. The disc is then alternately exposed to the direct solar rays and shielded from the rays. By accurately timing this procedure the intensity of the solar radiation can be determined. The instrument has always been calibrated against the primary standard pyrheliometer which is a water flow calorimetric type of instrument. The silver disc pyrheliometer is sensitive to all wavelengths, but only can be used on solar radiation.

Monteith net radiometer or balance meter

This balance meter is of the utmost simplicity in design and is being used for agricultural meteorological studies at Rothamsted Experimental Station in England. It is comprised of a glass ring with outside diameter roughly four inches and inside diameter three inches. The glass is approximately one-eighth inch thick. Many turns of constantan wire are wound on the ring, and the outer half of each turn is covered with paraffin wax by dipping the ring in melted wax and slowly turning it around its axis. Then the entire ring is immersed in a copper-plating solution which plates copper onto the inside half of each turn of constantan wire. The paraffin is then removed and one has a circle of thermojunctions on the top side of the ring and another circle of junctions on the under side. The entire ring is blackened and mounted on the end of a long arm. The wires can be connected to an electronic potentiometer recorder and the output calibrated in terms of the net flux of radiation on the two surfaces.

Stoll-Hardy radiometer

The basic idea involved in the design of this radiometer is very old, but the present version of the instrument is new and takes advantage of the availability of a fast, sensitive detector, the flake thermistor. Hardy (1934) had originally used

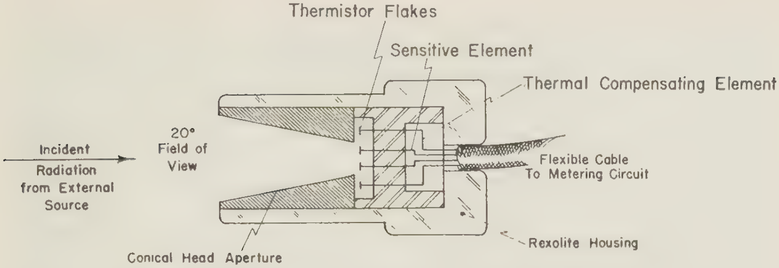

Thermistor Flakes
Sensitive Element
Thermal Compensating Element
20° Field of View
Incident Radiation from External Source
Flexible Cable To Metering Circuit
Rexolite Housing
Conical Head Aperture

FIGURE 22.—Stoll-Hardy radiometer for instantaneous observation of the radiant flux within a 20° field of view.

the device with a thermopile detector for physiological studies, and subsequently Stoll (1954) redesigned the instrument to make use of thermistor detectors. The design of the instrument as now used is shown in Figure 22. It is manufactured by the Williamson Development Corporation, West Concord, Massachusetts.

The radiometer measures the amount of radiant flux coming to it from within its field of view. It measures the temperature of a surface by virtue of the fact that all black objects radiate heat according to the law σT^4. For low temperature radiators most of the radiated energy will be at infrared wavelengths. If one is using the radiometer to measure a surface temperature which is illuminated by sunlight, an infrared filter of germanium must be placed across the entrance aperture to get a correct reading. It is usually best to use this filter under most conditions and to calibrate the instrument with the filter in position. The sensitive element consists of two thermistor flakes located at the apex of a polished aluminum cone having a 20° field of view. Two other thermistor flakes, matching the receiver pair, are located within the radiometer head unexposed to the incident radiation and act as temperature compensators. A microammeter is provided with three scales which permit readings departing from ambient of 0–10, 0–30, and 0–100 degrees centigrade. A cylindrical cavity drilled in an aluminum block housed within the instrument case provides the black body cavity against which all temperature readings are compared. The radiometer head is normally stored inserted into this cylindrical cavity and is only removed momentarily in order to take a reading of the radiation emitted by some surface. The temperature of the aluminum block must be carefully read by means of a mercury thermometer. Calibration of the radiometer may be achieved by means of a Leslie cube of unit emissivity into which may be poured water at various temperatures, water-ice mixtures, or other solutions. If the calibration data are plotted on an energy (σT^4) versus deflection graph, an excellent straight line should result. If temperature versus deflection is plotted, then a curve results which is more difficult and less accurate for extrapolation purposes.

Several modifications of the instrument are required to facilitate its use in the field. The detector head as provided by the company is not sufficiently well insulated to shield the thermistors from heat transferred from the hand through the head. If the head is encapsulated in a block of styrofoam or other insulating material, then its response and reproducibility are greatly improved. The aluminum reference block provided with the instrument has too small a heat capacity to remain nearly constant when used in the field. It is therefore desirable to provide a much more massive block of aluminum for this purpose. A cylinder six inches in diameter and six inches high, with a cylindrical cavity drilled in one end, proved suitable for this purpose when used under severe winter conditions. Wind can be quite disturbing to the detector, either by cooling or warming the head or the filter, or, if the aperture is not closed off by the filter, the wind enters the cone and disturbs the thermistors directly. Rather elaborate precautions can be taken by redesigning the entrance cone; however, this becomes a bit involved.

The primary disadvantage to the instrument in its present form is the fact that it is a d.c. device. This means that it is subject to any and all temperature disturbances of the types mentioned above. The instrument should be redesigned in order to convert it to an a.c. device in which the incoming stream of radiation to be detected is chopped. However, the thermistors currently used in the radiometer are too slow in response to permit this to be done. Several a.c. radiometers for radiant energy measurements in industry are currently on the market, but are generally much too expensive and not designed specifically for application to geophysical or biological problems. These are available from the Williamson Development Company, the Servo Corporation of America, and Barnes Engineering Corporation, to mention a few.

Despite the drawbacks to the Stoll-Hardy radiometer for field use, it is the only instrument currently available which can be adapted for field measurements where it is desired to obtain the temperatures of surfaces which are small, but greater than about 4 cm^2. The instrument may also be used to measure the radiant flux from the sky within the 20° field of view. However, by sampling the sky in several directions, it is even possible to make a good estimate of the total flux from the entire hemisphere incident upon a horizontal surface. The performance of the instrument in the field has been reported by Gates (1961) and by Stoll and Hardy (1955).

Frigorimeter

This instrument cannot be truly called a radiation measuring device since in fact it is designed to measure the complete heat transfer from the blackened sphere to its surroundings and to give an indication of the "physiological" cooling or heating power of the environment. The frigorimeter, a nearly solid copper sphere 7.5 cm in diameter, blackened on the outside and exposed to the atmosphere, is maintained at approximately 37°C by means of an electrical heater element in the center. All heat exchange factors, sunlight, thermal infrared radiation, wind, and precipitation affect the response of the instrument. The amount of current supplied is a measure of the cooling of the sphere. The instrument has undergone extensive testing and calibration at the Physikalisch-Meteorologischen Observatorium, Davos, Switzerland, and would appear to be satisfactory for the measurement of combined meteorological effects on the heat exchange of a blackened metal sphere. The extent to which measurements of this type can be interpreted in terms relative to living organisms, such as human comfort factors, may be subject to considerable question. However, the publications by Dorno (1926, 1928) discuss the matter in considerable detail. The construction details are given in an article by Thilenius (1931).

Nonventilated enclosed surface radiometers

These instruments are shielded from the free air by means of a dome or layer of polyethylene or mylar. These plastic materials when used in very thin pieces are transparent to most wavelengths including the infrared. Their principal disadvantage is that they are fragile and can be easily punctured; however they have been used successfully in the instruments described below.

Schulze radiometer

This radiometer is similar in appearance to the Moll-Gorczynski solarimeter, except here a polyethylene (Lupolen-H) hemispherical shield is used instead of the glass dome. Figure 23 (see opposite p. 79) shows the polyethylene dome removed and a small box containing two pairs of thermopiles. A single thermopile is made from constantan and silver wire with a total resistance of about 500 ohms and then is blackened with Parson's Matt Black Lacquer. Two thermopiles (one pair) are connected in parallel to give a net resistance of 250 ohms. In the earlier models first reported on by Schulze (1953) a Moll thermopile was used, but for the later models, such as the one shown in Figure 23, a thermopile of their own design was employed. The later design has been reported on by Fleischer (1958). The instrument has ample sensitivity for measuring accurately the long-wave radiation from the sky or the ground at night. The instrument is available commercially from B. Lange Company, Berlin, Germany. As shown in Figure 23, there are two radiometers, an upward facing and a downward facing, mounted back to back in order that the downward and the upward streams of radiation may be simultaneously measured. If the radiometers are connected in opposition, the net flux of radiation is measured. A thermocouple is located beneath each receiving surface in order that the receiver temperature can be measured so that back radiation by the receiver can be calculated. It is recommended that the cold junction for this thermocouple be buried in the ground at a depth of at least one meter.

The Schulze radiometer or balance meter is excellently designed and constructed. Two tubes seen in Figure 23 allow one to circulate dry air or nitrogen through the upper and lower hemispherical enclosures thereby drying them of any moisture which may otherwise condense within. The hemispherical dome can be easily removed for replacement, which should usually be done about once every two or three months. Any standard electrical recording potentiometers may be used with the instrument. Four channels are required for a single radiometer including the recording of the instrument temperature.

Suomi "economical" net radiometer or balance meter

In an attempt to produce an inexpensive, easily constructed net radiometer, Suomi (1958) reported on the design of such an instrument which subsequently has received a considerable amount of attention and use in meteorology and agriculture (see Suomi, Staley, and Kuhn, 1958).

The sensing element consists of a thermistor or thermometer sandwiched between two horizontal blackened foils of aluminum which are supported by a block of lightweight rigid material of low heat capacity and thermal conductivity such as fiber glass insulation. Two sheets of 0.50 mil thick polyethylene film, an insulating guard, and a wooden box framework form a wind screen. This whole laminated unit faces upward and a similar unit faces downward. Polyethylene is very transparent throughout infrared wavelengths and therefore is an excellent shield for instruments designed to detect long-wave radiation.

The "economical" radiometer has the following advantages: (*a*) The radiometer has high "thermal gain" so that easy temperature measurements result in reasonably accurate radiation estimates. (*b*) The instrument can be modified for broad spectral selection so that the upward and downward flux of total solar and thermal radiation can be measured in the daytime to ±5%. The instrument is also useful for measuring the nocturnal thermal radiation streams. The instrument can be fabricated very easily from materials which are readily available and at very low cost. The instrument is now being used from balloons as a radiation radiosonde to determine the radiation-divergence field of the atmosphere at various elevations. In addition, it has been used as a frost forecasting device by the cranberry growers of Wisconsin, and researchers in the United States Forest Service are experimenting with it for use within forested areas. Goodell (1962) has made this instrument into an automatic recording instrument by means of a digitized totalizer.

Funk net radiometer or balance meter

Upward facing and downward facing thermopiles of 250 junctions each are shielded from the external environment by hemispheres of very thin (0.002 inches) polyethylene which is coated with a thin film of silicone oil. Unfortunately the silicone oil reduces the transmission and introduces numerous weak absorption bands. The very thin polyethylene is employed, not so much to improve its transparency, but to avoid departure from cosine law behavior by the instrument. The very thin polyethylene cannot itself remain distended in the hemispherical shape, and so dry nitrogen is bled to the inside to maintain a slight pressure excess over atmospheric pressure. The originator of this instrument, Funk (1959, 1960), designed it to avoid the bulkiness of the Schulze radiometer and the inaccuracy of the Suomi "economical" net radiometer or balance meter, and at the same time to create an unaspirated instrument. The instrument appears to be very good for its intended use; however, it might not be sufficiently rugged for field use involving moderate to heavy wind, heavy precipitation, and hail. A special feature of the radiometer is a heating ring located equatorially which can radiate to both hemispheres, thereby warming them slightly to prevent the deposit of dew on the polyethylene surface. The radiometer has a time constant of about one minute and according to Funk has a sensitivity almost an order of magnitude higher than any previously reported net radiometers. The actual sensitivity is not given by Funk in his discussion however. The instrument is manufactured for sale by the C.S.I.R.O., Division of Meteorological Physics, Melbourne, Australia.

Fritschen and van Wijk net radiometer or balance meter

In an attempt to obtain a small, inexpensive, all-weather net radiometer, Fritschen and van Wijk (1959; Fritschen, 1960) designed an instrument similar in principle to the Suomi "economical" radiometer but utilizing as the sensing unit a small heat transducer, which was probably simply a thermopile rather than a thermometer. The thermal transducer is sandwiched between two sheets of mica or saran wrap. Two more surfaces of mica or saran wrap are stretched horizontally one-sixteenth inch above and below the first two surfaces. Both mica and saran wrap are less transparent to infrared radiation than is polyethylene; however, each of these is more impervious to water and therefore more suitable for field use. As far as could be ascertained by Fritschen, there was no noticeable bias in the response of the mica or saran wrap protected units compared with those covered with polyethylene.

Chapter 4

Heat Transfer by Convection

INTRODUCTION

It will be recalled that an organism immersed in a medium —air, water, or soil—will undergo changes in energy by the processes of radiation, convection, conduction, heat transfer, and evaporation or condensation. For objects, such as roots or organisms in the soil, the heat flow will be primarily by conduction. For objects in water or in air, the conduction and convection processes will share in the transfer of energy. The process of conduction transfers heat from a region of higher temperature to a region of lower temperature by direct molecular activity through the medium, without mass motion of the medium itself taking place. The process of convection transfers energy by the combined effect of conduction and mass transport through the fluid. In convection, heat is transferred from the surface of the organism to the fluid by conduction, but within the fluid itself it is carried along by the movement of a stream of fluid which transports molecules of higher energy content from points of higher to points of lower temperature. The effectiveness of energy transfer by convection depends largely upon the mixing action of the fluid.

In the process of heat transfer by convection there are two classes or modes of action: (*a*) free or natural convection in which the flow is created solely because of density differences caused by temperature differences within the fluid; (*b*) forced convection in which the flow is caused by an external force field such as a blower or some sort of pressure difference within the fluid creating a "wind." In order to understand the process of convection, one must first understand the detailed physical laws occurring within the fluid.

At the surface of the object or organism, across which the fluid (air or water) may be flowing, the fluid must be stationary since clearly the surface cannot be flowing relative to itself. The full velocity of the fluid is achieved as one moves away from the surface. The transition zone, within which the velocity is increasing as one moves outward, is referred to as the "boundary layer." This is illustrated in Figure 24. The boundary layer may be only a fraction of a millimeter thick. All objects within a moving fluid possess this boundary layer to some degree. It represents a transition zone for fluid motion and for temperature difference between the object and the fluid. It is demonstrated most dramatically for the human form when one is taking a hot bath such as the Finnish sauna. The sauna, a small enclosed room in which the air is moist but of low relative humidity, is often at temperatures as high

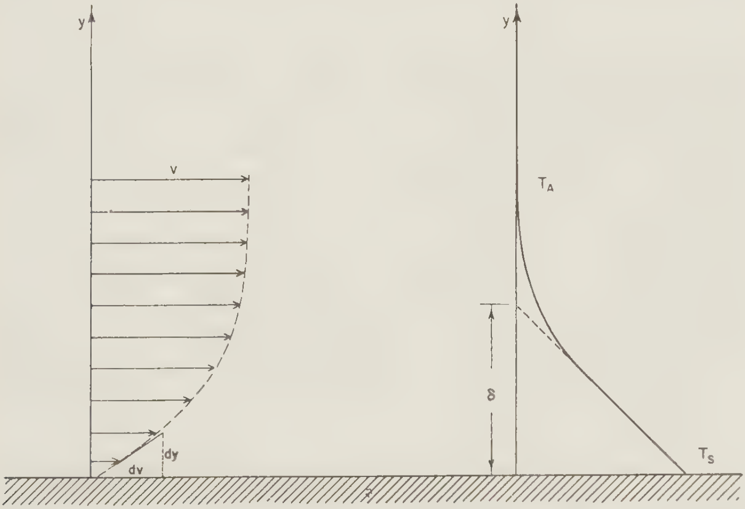


Figure 24.—Wind velocity and air temperature profiles near a plane surface when the wind speed in the free air is v, the air temperature T_A, and the surface temperature T_S. The boundary layer thickness next to the surface, δ, is defined in terms of the temperature profile. A change in wind speed, dv, exists for a change in height, dy.

as 230°F. This is truly a "black body" cavity filled with infrared radiation. The skin obviously cannot attain a temperature of 230°F without burning, and indeed it does not reach this temperature, but a much lower temperature, because of the boundary layer of stationary air next to the skin. If, while in this situation in the sauna you blow on your skin, you destroy this boundary layer, and immediately a burning sensation is experienced; indeed a burn can result.

The process described as the "drag" of a fluid when flowing across the surface of an object is a result of its viscosity. The force required to move the fluid across the surface is proportional to the area of contact between the surface and the fluid; the greater the area of contact, the greater the force required. The force is also proportional to the velocity gradient at the interface; the greater the force exerted, the greater the velocity flow at some point near the surface. These relationships can be represented in the following form:

$$f = \mu A \left.\frac{dv}{dy}\right|_{y=0} \tag{4.1}$$

where f represents the force in dynes, A the area in cm^2, dv/dy the velocity gradient with v the velocity in cm per sec and y the distance perpendicular to the surface in cm. The symbol dv represents a small change in velocity associated with a small change dy in distance from the surface as illustrated in Figure 24. The proportionality constant, μ, is called the viscosity coefficient and is given in units of grams per sec-cm or poises, named after Poiseuille who was one of the first to study viscous flow. Table VII gives the viscosity coefficient for water and air as a function of the temperature in °C. The viscosity coefficient will enter into our discussion of energy transfer by convection further on.

Heat is energy in transition. It is transferred from the surface of the object or organism by conduction in the boundary layer near the surface and then is carried away by convection through the fluid. Clearly the amount of energy transferred to or from the surface by conduction must be equal to the amount of energy transferred to or from the outer region of

TABLE VII

Typical Values of Viscosity for Water and Air (in Poises)

Temperature (°C)	Water	Air
0	1.792×10^{-2}	1.71×10^{-4}
20	1.005	1.81
40	0.656	1.90
60	0.469	2.00
80	0.357	2.09
100	0.284	2.18

the boundary layer by convective motion in the fluid. If the two were not equal, there would be a net heating or a net cooling within the boundary layer, an event which does not occur for the steady state situation. The amount of energy or heat, dQ, in calories transferred in a unit of time, dt, in seconds is proportional to the temperature difference, ΔT, in °C between the surface and the air ($\Delta T = T_s - T_a$), and is proportional to the area, A, of the surface in cm^2 in contact with the fluid flow. This proportionality can be written as an equation by means of a proportionality constant, h_c, called the convection coefficient.

$$\frac{dQ}{dt} = h_c A \Delta T \tag{4.2}$$

The units for h_c will be cal per cm^2 per sec per °C. The problem of determining h_c is often a very difficult one, since it involves a number of factors describing the geometry and physics of the particular problem. The convection coefficient depends upon:

1. Shape of surface—flat or curved.
2. Orientation of surface—horizontal or vertical.
3. Physical properties of the fluid—density, viscosity, specific heat, and thermal conductivity.
4. Whether evaporation or condensation takes place.
5. Whether the velocity of the fluid is small enough to give

rise to laminar flow, or large enough to give rise to turbulent flow.

In the boundary layer, if heat is transferred only by conduction, one can say that the rate of flow of heat is proportional to the surface area, A, in contact with the fluid and proportional to the temperature gradient, dT/dy, at the surface. An equality can be written using a proportionality constant, k, in this case called the coefficient of thermal conductivity given in cal per sec per cm per °C.

$$\frac{dQ}{dt} = -kA\frac{dT}{dy} \tag{4.3}$$

The minus sign is used in order that the positive direction of the flow of heat should coincide with the positive direction for y, which is positive upward. The thermal conductivity for air is about 6.0×10^{-5} cal per sec per cm per °C at 20°C and for water is 1.43×10^{-3} cal per sec per cm per °C.

As stated earlier, the heat conducted through the boundary layer must equal the heat to be carried away by convection. The temperature gradient can be approximated by the temperature difference across the boundary layer divided by the thickness of the boundary layer, δ; hence

$$\frac{dT}{dy} = \frac{T_s - T_a}{\delta}$$

Therefore, equating Equations 4.2 and 4.3

$$\frac{dQ}{dt} = kA\frac{(T_s - T_a)}{\delta} = h_cA(T_s - T_a) \tag{4.4}$$

we get

$$h_c = \frac{k}{\delta} \tag{4.5}$$

Certain dimensionless variables may be formed relating the basic parameters of the media and the dimension of the object. Heat transfer engineers find these useful for there are certain invariant relationships between the dimensionless variables, and once these relationships are established for a given object, they then can be evaluated for any media. In the case of natural or free convection, there are three dimensionless variables. They are as follows:

$$\text{Nusselt number, Nu} \equiv \frac{h_c L}{k} = \frac{L}{\delta} \tag{4.6}$$

where L is a significant or characteristic dimension of the object concerned. The characteristic dimension would be the diameter of a sphere or cylinder, and the length of a plane. Clearly the Nusselt number is a measure of the ratio of the characteristic dimension to the boundary layer thickness.

$$\text{Prandtl number, Pr} \equiv \frac{c_p \mu}{k} \tag{4.7}$$

where c_p is the specific heat of the gas at constant pressure, μ is the viscosity, and k the conductivity. The Prandtl number measures the relative efficiency of the conducting system for the molecular transport of momentum and energy.

$$\text{Grashof number, Gr} \equiv \frac{\rho^2 g \beta (T_s - T_a) L^3}{\mu^2} \tag{4.8}$$

where ρ is the density of the fluid, g the acceleration due to gravity, β the coefficient of expansion, T_s and T_a the temperature of the surface and the air, respectively, L the characteristic dimension, and μ the viscosity.

FREE OR NATURAL CONVECTION

In the case of free or natural convection it can be shown through dimensional analysis and other arguments that the three dimensionless variables are related in the following manner.

$$\text{Nu} = \text{f}(\text{Gr}, \text{Pr}) \tag{4.9}$$

For natural or free convection in which the temperature difference between the object and the surrounding medium creates a flow of air, if the medium is air, the heat transfer engineers have arrived at values of the convection coefficient for objects of various shapes, sizes, and orientations. Two cases are of interest from the standpoint of relationships to biological organisms: flat plates, representing an approximation to a leaf, and pipes or cylinders representing all cylindrically shaped objects in nature such as tree branches, twigs, needles, and the bodies of most animals. The following illustrations will show the method for proceeding to the Nusselt number and then to the convection coefficient. Table VIII contains the Nusselt numbers for various situations which are applicable to biological problems.

For air the value of the Prandtl number is 0.72 and the Grashof number is 156 ΔTL^3 at 20°C. The Nusselt number for a heated plate facing upward in air at 20°C, for the laminar range, will be

$$\mathrm{Nu} = 0.71\ (156\ \Delta TL^3\ 0.72)^{1/4} \tag{4.10}$$

From this one gets for the convection coefficient, in the laminar range from Equation 4.6,

$$h_c = 1.31 \times 10^{-4} \left(\frac{\Delta T}{L}\right)^{1/4} \tag{4.11}$$

TABLE VIII

Nusselt Numbers for Free or Natural Convection (in terms of the Grashof number, Gr, and the Prandtl number, Pr)

1. Horizontal cylinders	$\mathrm{Nu}^{1/2} = 0.63 + 0.35(\mathrm{Gr} \times \mathrm{Pr})^{1/6}$
2. Vertical cylinders	
Laminar range	$\mathrm{Nu} = 0.726(\mathrm{Gr} \times \mathrm{Pr})^{1/4}$
Turbulent range	$\mathrm{Nu} = 0.067(\mathrm{Gr} \times \mathrm{Pr}^{1.29})^{1/3}$
3. Plane surfaces	
Laminar range	
Horizontal facing upward	$\mathrm{Nu} = 0.71(\mathrm{Gr} \times \mathrm{Pr})^{1/4}$
Horizontal facing downward	$\mathrm{Nu} = 0.35(\mathrm{Gr} \times \mathrm{Pr})^{1/4}$
Turbulent range	
Horizontal facing upward	$\mathrm{Nu} = 0.17(\mathrm{Gr} \times \mathrm{Pr})^{1/3}$
Horizontal facing downward	$\mathrm{Nu} = 0.08(\mathrm{Gr} \times \mathrm{Pr})^{1/3}$
Vertical	$\mathrm{Nu} = 0.13(\mathrm{Gr} \times \mathrm{Pr})^{1/3}$

The Nusselt number for a heated plate facing upward in air at 20°C, for the turbulent range, will be

$$\mathrm{Nu} = 0.82\,(\Delta T L^3)^{1/3} L \tag{4.12}$$

From this one gets for the convection coefficient, in the turbulent range from Equation 4.6,

$$h_c = 0.465 \times 10^{-4} (\Delta T)^{1/3} \tag{4.13}$$

The convection coefficients, h_c, for laminar flow are summarized in Table IX.

The surface of a cold horizontal plate facing downward has the same convection coefficient as the surface of a warm horizontal plate facing upward, and a cold upward facing surface has the same coefficient as a warm downward facing plate. A numerical example to illustrate the strength of heat transfer by natural convection is as follows. Consider a warm horizontal plate of dimension 1 cm, facing upward with a surface temperature 10°C above the air temperature, which is a reasonable approximation to conditions encountered by the leaves of vegetation. To calculate the quantity of heat lost per unit time per unit area by natural convection, one uses Equation 4.2 and the value given in Table IX.

$$\begin{aligned} h_c &= 7.86 \times 10^{-3} (\Delta T)^{1/4} = 7.86 \times 10^{-3} (10°\mathrm{C})^{1/4} \\ &= 14.04 \times 10^{-3} \text{ cal per cm}^2 \text{ per sec per °C} \end{aligned}$$

Then

$$\frac{1}{A}\frac{dQ}{dt} = h_c \Delta T = 14.04 \times 10^{-3} \times 10°\mathrm{C} = 0.140 \text{ cal per cm}^2 \text{ per min}$$

TABLE IX

CONVECTION COEFFICIENTS (IN CAL PER CM² PER SEC PER °C) FOR FREE OR NATURAL CONVECTION IN LAMINAR FLOW (ΔT IS THE TEMPERATURE DIFFERENCE IN °C BETWEEN THE SURFACE OF THE OBJECT AND THE SURROUNDING AIR. L IS THE DIMENSION OF THE PLATE IN THE DIRECTION OF FLOW.)

Warm horizontal plate, facing upward	$7.860 \times 10^{-3} (\Delta T/L)^{1/4}$
Warm horizontal plate, facing downward	$3.866 \times 10^{-3} (\Delta T/L)^{1/4}$
Vertical plate	$6.042 \times 10^{-3} (\Delta T/L)^{1/4}$
Horizontal or vertical pipe of diameter D	$6.000 \times 10^{-3} (\Delta T/D)^{1/4}$

It is seen that the amount of energy transferred by natural convection from a horizontal leaf whose surface temperature is 10°C above the surrounding air is rather small and may almost be neglected compared to the energy transferred by radiation, transpiration, and wind. This does not imply, however, that one should neglect understanding the role of natural convection among the processes affecting the energy exchange of a leaf. Furthermore, as the heat transfer by natural convection for small ΔT becomes small, there still remains a residual rate of heat transfer by means of straight conduction which depends upon the natural conductivity of the air. This can be estimated from Equation 4.3 as follows:

$$\frac{1}{A}\frac{dQ}{dt} = -36.0 \times 10^{-4}\frac{dT}{dy} \text{ cal per cm}^2 \text{ per min}$$

If the boundary layer was 4 mm thick across which the temperature difference between the leaf surface and the air was 10°C, then

$$\frac{1}{A}\frac{dQ}{dt} = 0.009 \text{ cal per cm}^2 \text{ per min}$$

If the boundary layer was only 2 mm thick the amount of heat transferred by conduction alone would be 0.018 cal per cm^2 per min. It is doubtful if conditions exist which could make this much greater.

FORCED CONVECTION

The heat transfer relationships for the case of forced convection must clearly involve the wind effect or the velocity of the medium relative to the object. The flow of air across a surface can most conveniently be defined by means of a dimensionless variable called the Reynolds number which is defined as follows:

$$\text{Re} \equiv \frac{\rho v L}{\mu} = \frac{vL}{\eta} \tag{4.14}$$

where ρ is the density of the fluid, v the velocity of the flow, L

the dimension of the object, μ the coefficient of viscosity, and η the coefficient of kinematic viscosity which is simply the ratio of μ/ρ. Reynold's formulation of the interaction of the fluid flow with surfaces was originally applied to pipes in which L was the diameter of the pipe. The formulation has now been extended to describe the properties of flow over surfaces such as the smooth earth where the following implications exist:

1. Laminar flow exists within a depth δ, called the laminar sublayer of the smooth surface when the following condition holds:

$$\frac{v\delta}{\eta} \cong 55 \tag{4.15}$$

If $v = 447$ cm per sec (10 mph) and $\eta = 0.14$ cm^2 per sec for air, $\delta = 0.017$ cm $= 0.17$ mm. Therefore laminar flow exists only within a very thin boundary layer.

2. Turbulent flow exists above a height δ' given by:

$$\frac{v\delta'}{\eta} \cong 750 \tag{4.16}$$

If $v = 447$ cm per sec (10 mph) and $\eta = 0.14$ cm^2 per sec for air, $\delta' = 0.235$ cm $= 2.35$ mm. At heights above the surface between δ and δ' one is in the transition region. The roughness characteristics of a surface and its influence on the flow of air will be discussed in more detail later on.

For the situation of forced convection it is found that the relationship between the dimensionless variables is as follows:

$$\mathrm{Nu} = f(\mathrm{Gr}, \mathrm{Pr}, \mathrm{Re}) \tag{4.17}$$

However, the Grashof number only accounts for the heat transferred by the natural convective "chimney" effect and is in general negligible compared to the forced current of air. Of course there will be an intermediate situation in which the forced current is weak and the Grashof number must be included. This situation may be one of common occurrence in ecology. With the Grashof number ignored, the functional relationship should be:

$$\mathrm{Nu} = f(\mathrm{Pr}, \mathrm{Re}) \tag{4.18}$$

For the case of forced convection in laminar flow over plane surfaces, the distinction between the upward facing and the downward facing surface no longer is important, and the Nusselt number is given by:

$$\mathrm{Nu} = 0.664\ \mathrm{Re}^{1/2}\mathrm{Pr}^{1/3} = 0.595\ \mathrm{Re}^{1/2} \tag{4.19}$$

The Reynolds number can be calculated from Equation 4.14 and the values are given in Table X for various wind speeds and plate sizes (the characteristic length).

Using a Prandtl number for air of 0.72, one can now calculate for each Reynolds number a corresponding Nusselt number and an appropriate convection coefficient for each characteristic length. These are given in Table X. If one were to write out the equation for the convection coefficient it would have the following form:

$$\begin{aligned} h_c &= \mathrm{Nu}\,\frac{k}{L} = 0.664\left(\frac{vL}{\eta}\right)^{1/2} \mathrm{Pr}^{1/3}\,\frac{k}{L} = 0.664\,\frac{k}{\eta^{1/2}}\,\mathrm{Pr}^{1/2}\,\frac{v^{1/2}}{L^{1/2}} \\ &= 9.55 \times 10^{-5} \sqrt{\frac{v}{L}}\ \text{cal per 1 cm}^2\ \text{per sec per °C} \qquad (4.20) \\ &= 5.73 \times 10^{-3} \sqrt{\frac{v}{L}}\ \text{cal per cm}^2\ \text{per min per °C} \end{aligned}$$

By means of Equation 4.4 and the convection coefficients given in Table X, one can calculate the rate of heat transfer per unit area for various conditions of temperature difference, plate size, and wind speed. These are given in Table XI in cal per cm^2 per min.

Several interesting features concerning the heat transfer rates for plane surfaces, which in nature would be the leaves of deciduous plants, may now be commented on. The smaller the surface dimension along the direction of wind flow, the greater will be the rate of energy transfer. A plant with many small leaves will lose energy due to forced convection much more rapidly than a plant with a few large leaves. This is because the air flowing across the leaf has its greatest cooling influence near the leading edge and progressively less effect as it moves away from the leading edge of the leaf. The heat

TABLE X

REYNOLDS AND NUSSELT NUMBERS AND CONVECTION COEFFICIENTS (IN CAL PER CM^2 PER MIN PER °C) FOR FORCED CONVECTION ACROSS A FLAT PLATE AS A FUNCTION OF WIND SPEED AND PLATE DIMENSION

L		v		
		5 mph (223 cm/sec)	10 mph (447 cm/sec)	30 mph (1341 cm/sec)
	Re	1596	3192	9576
1 cm	Nu	23.8	33.6	58.3
	h_c	0.085	0.120	0.208
	Re	7980	15950	47880
5 cm	Nu	53.2	75.1	130.3
	h_c	0.038	0.054	0.093
	Re	15960	31920	95760
10 cm	Nu	75.3	106.5	184.6
	h_c	0.027	0.038	0.066

transfer rate increases only with the square root of the wind speed and not linearly with it. A 20 mph wind will produce only twice the energy transfer rate as a 5 mph wind in laminar flow. The heat transfer rate increases linearly with the temperature difference between the leaf surface and the surrounding air, whereas for the case of natural convection the temperature difference entered only to the fourth root, a much weaker influence. The amount of energy lost or gained by a leaf through forced convection may be substantial. If the heat load on one surface of a leaf from sunlight and thermal radiation was 2.0 cal per cm^2 per min and the leaf temperature reached 5°C above the ambient air temperature with a wind speed of 10 mph, the convective heat loss would carry away 0.600 cal per cm^2 per min for a leaf of 1 cm dimension, or account for nearly one-third of the total energy transferred. Leaf temperatures as high as 10°C above ambient air temperature frequently exist and for this, with a 10 mph wind speed and a leaf size of 1 cm, the convective transfer rate would be 1.200 cal per cm^2 per min. These numbers, of course, assume that a

TABLE XI

Rate of Heat Transfer (per unit area in cal per cm^2 per min) for Forced Convection across a Flat Plate as a Function of the Temperature Difference between Surface and Air, the Characteristic Dimension of the Surface, and the Wind Speed

L	ΔT				
	1°C	5°C	10°C	20°C	40°C
		v = 5 mph = 223 cm/sec			
1 cm	0.085	0.425	0.850	1.600	3.200
5 cm	0.038	0.190	0.380	0.760	1.520
10 cm	0.027	0.135	0.270	0.540	1.080
		v = 10 mph = 447 cm/sec			
1 cm	0.120	0.600	1.200	2.400	4.800
5 cm	0.054	0.270	0.540	1.080	2.160
10 cm	0.038	0.190	0.380	0.760	1.520
		v = 30 mph $\doteq$ 1341 cm/sec			
1 cm	0.208	1.040	2.080	4.160	8.320
5 cm	0.093	0.465	0.930	1.860	3.720
10 cm	0.066	0.330	0.660	1.320	2.640

leaf, as far as the energy transfer by convection is concerned, behaves very much the same as an inanimate flat object. Apparently this is the case to a reasonable approximation according to the excellent work of Raschke (1960) on this subject. Raschke says that the heat transfer coefficient has the following dependence upon leaf dimension and wind speed:

$$h = (\text{constant})\ (L)^{-0.3}(v)^{0.5} \tag{4.21}$$

The constant depends upon the shape of the leaf, which we know will influence the heat transfer rate. Few leaves are perfectly flat planes but are curled slightly. This will tend to reduce the efficiency of the heat transfer mechanism by convection. The formula given by Raschke is nearly the same as Equation 4.20, which was used for computing Table X. They differ slightly according to the exponent for the characteristic length, L, which is the dimension of the leaf in the direction of wind flow. Despite this small difference, it does demonstrate that good approximation to the natural situation for leaves can be achieved using strictly physical arguments and the standard deductions of heat transfer theory.

Many objects in nature, such as limbs, branches, needles, the bodies of most animals, can be approximated by a cylinder. For this reason it is important to acquire an understanding of the heat transfer from a cylinder. Theory shows that for forced convection across a cylinder the following relationship should hold in air:

$$h_c = \mathrm{Nu}\,\frac{k}{D} = 0.89\left(\frac{vD}{\eta}\right)^{1/3}\frac{k}{D} = 6.17 \times 10^{-3}\,\frac{v^{1/3}}{D^{2/3}} \qquad (4.22)$$

where h_c will be in cal per cm^2 per min per °C.

Table XII gives some numerical values for the rate of heat transfer from cylinders in a wind stream of air at 10 mph. It is seen that very small objects may have fairly large heat transfer rates, although for pine needles, as an example, it is doubtful if their temperature gets very much above ambient air temperature.

It is evident that the flow of air over the surface of the earth has important ecological implications, and conversely the type of growth on the surface manifestly affects the passage of the air, creating a sheltered air layer close to the surface and disturbing the flow to a considerable height above the vegeta-

TABLE XII

RATE OF HEAT TRANSFER (IN CAL PER CM^2 PER MIN) FOR FORCED CONVECTION ACROSS A CYLINDER AS A FUNCTION OF TEMPERATURE DIFFERENCE AND CYLINDER DIAMETER (v = 10 MPH = 447 CM PER SEC)

D	ΔT				
	1°C	5°C	10°C	20°C	40°C
0.1 cm	0.219	1.095	2.190	4.380	8.760
1.0 cm	0.047	0.235	0.472	0.944	1.888
5.0 cm	0.016	0.080	0.161	0.322	0.644
10.0 cm	0.010	0.051	0.102	0.204	0.408

tion. The daytime circulation pattern is generally a combination of forced eddy-convection and natural convection produced by the overhead winds and the solar heating of the surface. The nocturnal situation is generally one of stable stratification through radiational cooling and the subsidence of cooler air downhill into depressions. In each case the flow of air is affected by the frictional drag force, τ_o, of the surface layer. From the surface upward the type of flow may be described as a laminar sublayer close to the surface, then the surface layer extending to a height of 20 meters or so above the surface, up to approximately 1000 meters, and finally the free air above all of this. All the layers below the free air region are considered a part of the friction layer characterized by the degree of surface roughness.

The wind profile in the surface layer above vegetation can be described in the following form:

$$\bar{u} = \frac{u_*}{k} \, ln \, \frac{z - d}{z_o} = \frac{1}{k} \left(\frac{\tau_o}{\rho}\right)^{1/2} ln \left(\frac{z - d}{z_o}\right) \qquad (4.23)$$

where $\bar{u}$ is the mean wind speed in the surface layer; u_* is the friction velocity and as a rule of thumb can be considered to be a velocity of about one-tenth the velocity at 2 meters above the surface; z is the height above the surface; z_o is the roughness parameter, which varies with the nature of the terrain and the type of vegetation; d is a displacement of the equivalent zero plane; k is a constant having the value of 0.40; and τ_o is the drag force on the surface. The relationships expressed by Equation 4.23 are shown in Figure 25. If the wind velocity followed a linear relationship with height, then only z_o would be required to describe the intercept. However, because of the influence of the vegetation or the roughness of the surface the curve becomes concave upward as the equivalent zero plane is approached; d is required to describe this property. Values of the roughness parameter, z_o, are as follows: sand, 0.01 to 0.1; snow surface, 0.1 to 0.6; short grass, 0.6 to 4.0; long grass, 4.0 to 10.0; and taller vegetation will have values in the range greater than 10.0. The drag force will vary between 1.0 and 10.0 dynes per cm² at the earth's surface.

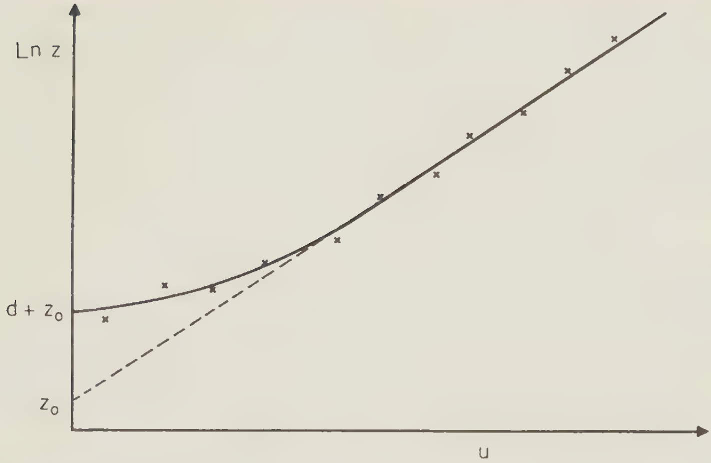


FIGURE 25.—The wind velocity profile above a very rough surface. u is the wind velocity; z is the height above the surface; z_0, the roughness parameter, is the extrapolated zero height if the wind velocity followed a linear relationship; and d is the displacement of the equivalent zero plane.

Over a forested area there will be large-scale eddies which will extend out into a clearing for a considerable distance with strong downward eddies developing downwind from the edge of the trees. These eddies will decay in intensity downwind, but will merge with small-scale eddies developing over the grass of the open meadow. The true wind flow typical of a broad expanse of meadow will only develop if the clearing in a pine forest is at least one-half mile in diameter. Meteorological instruments located at the center of such a clearing can be considered to be in an open exposure.

Since the flow of air above any vegetated surface can be characterized by means of eddies, a considerable amount of attention has been centered on the problem of the eddy diffusion of momentum, water vapor, and heat near the ground. The entire problem of the eddy transport of these quantities from vegetation into the surrounding air is important to ecology; however, it is also a difficult and fairly complex subject. The formulation of the problem will just be stated here without going into a detailed discussion of the application to particular situations. Rider (1954) and Rider and Robinson (1951) have shown that in the first two meters above a short grass surface, the profiles of temperature, momentum (wind), and humidity are nearly identical. The following equations describe the eddy diffusion of heat, moisture, and momentum or wind.

1. Eddy transfer of heat

$$C(z) = -K(z)\,\rho\, c_p \left(\frac{\partial T}{\partial z}\right) \tag{4.24}$$

where $C(z)$ is the nonradiative flux of heat at height z in cal per cm^2 per sec and $K(z)$ is a proportionality factor termed the eddy conductivity in cm^2 per sec.

2. Eddy transfer of moisture

$$E(z) = -D(z)\,\rho \left(\frac{\partial e}{\partial z}\right) \tag{4.25}$$

where $E(z)$ is the vertical flux of water vapor at height z in gm per cm^2 per sec, e is the water vapor concentration in grams per gram of air, and $D(z)$ is a proportionality factor termed the eddy diffusivity of water vapor in cm^2 per sec.

3. Eddy transfer of momentum

$$M(z) = -N(z)\,\rho \left(\frac{\partial u}{\partial z}\right) \tag{4.26}$$

where $M(z)$ is the vertical flux of momentum (mass multiplied by velocity in the vertical direction) at height z in gm per cm per sec^2, u is the wind speed in cm per sec, and $N(z)$ is a proportionality factor termed the eddy viscosity in cm^2 per sec.

Equations 4.24, 4.25, and 4.26 can be evaluated if one measures the temperature, moisture, and wind velocity at two or more heights above the vegetation, thereby determining the gradients of these quantities, namely, $\partial T/\partial z$, $\partial e/\partial z$, and $\partial u/\partial z$, and if one makes certain assumptions. The assumptions are that the three coefficients, N, D, and K, are equal.

Furthermore, from Equation 4.23, if one identifies τ_0 with $M(z)$ in the steady state, one gets from Equation 4.26

$$N(z) = k^2 z^2 \frac{\partial u}{\partial z} \tag{4.27}$$

Rider and Robinson (1951) and Rider (1954) have evaluated these quantities in detail and for further information the reader must refer to the original articles.

The assumption that the turbulent transfer for heat, water vapor, and momentum is the same has been challenged from time to time, but Raschke (1955) has more recently reconfirmed the assumption. His first experiment was done in the laboratory by measuring the gradients of moisture and temperature near the surface of wet filter paper. His second experiment was done with a leaf in its natural environment outdoors and again confirmed the equality for the coefficients of heat and water vapor transfer. In addition he attempted to measure the difference between the heat transfer by convection from the upper surface of a leaf and from the lower surface of a leaf for nearly still air and came to the conclusion that they were equal. This contradicts the theory that has been so well confirmed by heat transfer engineers, which states that for natural convection the energy lost from the upward facing surface of a horizontal heated plate is nearly double that lost from the downward facing surface. More measurement should be made in order to give accurate values for leaf surfaces. However, for the small temperature differences existing between the leaf surface and the surrounding air, the energy loss through natural convection is small compared to the loss by radiation and transpiration.

Since space is limited, this discussion of the eddy diffusion has been introduced mainly to point out that in making ecological studies more quantitative, one must deal with the true situation. It is not correct to measure the wind speed within a certain ecosystem and then expect to relate its influence on the vegetation directly to some property of the vegetation. Rather, one must couple the effects of the wind with the vegetation through such relationships as those given for the eddy diffusion and the theory of heat transfer by convection. Similarly one should not just measure the moisture content of the air, but rather one should measure moisture gradients and by means of relationships, such as Equation 4.25 for the eddy transfer of moisture, determine the actual exchange rate of moisture. Measurements of the moisture profile near the surface of a leaf were made by Ramsay, Butler, and Sang (1938) and by Raschke (1954). The ecologist has far too long measured certain climatological factors which are interesting features of the environment, but has not related these parameters to physical events affecting the organism, such physical events being the transfer of moisture, heat, and gases such as carbon dioxide. Studies such as those discussed above must now be applied in detail to every sort of ecosystem in order to truly and correctly understand the interaction of the plants and animals with the environment.

Chapter 5

New Methods in Ecology

INTRODUCTION

In order to better determine the interaction of plants and animals with their environment, many new instrumentation techniques developed in other disciplines can now be applied to the field of ecology. It has been made clear in the earlier chapters that the ecologist must have a thorough knowledge of micrometeorology, heat transfer physics, some aspects of atmospheric physics, and certain mathematical techniques. A proper understanding of these subjects by the ecologist will give him a fertile background of knowledge and skills which will lead to many new ideas, techniques, and methods of use in ecology. A few possible suggestions of such techniques are given here.

MEASUREMENT OF CO_2 AND H_2O CONTENT OF THE ATMOSPHERE

A considerable need exists to determine on a continuous basis the carbon dioxide and water vapor content of the air within and above a stand of vegetation. Measurements of these quantities which are made at a single position or even at several specific positions are likely to be unsatisfactory in that they represent measurements only at these sites and may not be representative of the average conditions existing within or above a stand of vegetation. The absorption produced along a beam of light by specific gases distributed along a given line of sight will give spatial averages of the gas concentration along the path length. Use may be made of the infrared absorption bands due to carbon dioxide, water vapor, and ozone (shown in Figures 3 and 6) for measuring the quantities of these constituents along a given light path. Since the various absorption bands of water vapor or carbon dioxide have different strengths, it is feasible to determine the amounts of these gases present over a considerable range in concentration and path length.

The strengths of the near infrared absorption bands of water vapor and carbon dioxide have been carefully determined from laboratory measurements made at Ohio State University by Howard, Birch, and Williams (1956). The gas to be studied was placed in an absorption tube 22 meters long with mirrors mounted at either end. By adjustment of the mirrors, path lengths up to 100 times the tube length could be achieved. The gases were placed in the tube at different concentrations and pressures in order that the effect on the absorption of both of these parameters could be studied. The pressure effect is particularly important for consideration of the absorption along slant paths in the atmosphere where absorption along the path is taking place at different pressure levels. One end of the multipath absorption tube contained an intense source of infrared radiation such as a Nernst filament, globar, or 1000-watt projection lamp, and the other end of the tube focused the energy into an infrared prism spectrometer. The spectrometer could be set to scan across the particular band to be measured. Although the spectral resolution achieved by the spectrometer was very low and many spectral details are lost from the spectra, nevertheless the information obtained is very valuable from the standpoint of the behavior of the total band absorption as a function of the pressure, concentration, and path length. The fractional absorption at a given wavelength or frequency position is described by A_ν. A given band will extend between two frequency limits ν_1 and ν_2. The total absorption in the band will be given by the sum or integral of all the individual fractional absorptions across the band. This will be written as

$$\int_{\nu_1}^{\nu_2} A_\nu d\nu$$

The mean fractional absorption across the band will be given by:

$$\bar{A} = \frac{\int_{\nu_1}^{\nu_2} A_\nu d\nu}{\nu_1 - \nu_2} \qquad (5.1)$$

The investigators have fitted empirical formulas to the functional behavior of the mean fractional absorption with pressure and quantity of absorbing gas in the path. For each of the CO_2 and H_2O bands, these formulas are listed below and the frequency spread of the band is also given. It turns out that two formulas rather than one are required to fit most of the bands to a wide range of absorption values. One formula, the so-called "strong fit," will apply when the absorption is greater than a given amount; the other formula, the "weak fit," will apply when the absorption is less than the stated amount.

The frequency of the band limits will be given in wave numbers, written cm^{-1}. A wave number in cm^{-1} is the direct reciprocal of the wavelength expressed in cm. As an example, a wavelength of 10 microns $= 10^{-3}$ cm, and the reciprocal is 10^3 $cm^{-1} = 1000$ cm^{-1}. The reason for the reciprocal relationship is that the wavelength and the frequency of a wave can be expressed as $\lambda = c/\nu$ where λ is the wavelength in units of length such as cm, c the speed of the wave in cm per sec, and ν the frequency in vibrations per second. However, for light waves, c is a constant, the speed of light, and since λ and ν are then inversely proportional to each other, it is convenient to simply express the frequency in wave numbers as the direct inverse of the wavelength.

The formulas for the total band absorption in some of the more important CO_2 and H_2O infrared absorption bands are given below. In these, w is the amount of absorbing gas expressed in atmosphere-cm for CO_2 and in cm of precipitable water vapor for H_2O, P is the total pressure, and p the partial pressure of the absorbing gas, both expressed in mm of Hg.

15 micron CO_2 band $\nu_1 = 800$ cm, $\nu_2 = 550$ cm^{-1}

$\int A_\nu d\nu > 50$ cm^{-1}

$$\int A_\nu d\nu = -68 + 55 \log w + 47 \log (P + p) \quad (5.2)$$

$\int A_\nu d\nu < 50$ cm^{-1}

$$\int A_\nu d\nu = 3.16\, w^{1/2} (P + p)^{0.44} \quad (5.3)$$

5.2 micron CO_2 band $\nu_1 = 1870$ cm^{-1}, $\nu_2 = 1980$ cm^{-1}

$$\int A_\nu d\nu = 0.024\, w^{1/2} (P + p)^{0.40} \quad (5.4)$$

4.8 micron CO_2 band $\nu_1 = 1980$ cm^{-1}, $\nu_2 = 2160$ cm^{-1}

$$\int A_\nu d\nu = 0.12\, w^{1/2} (P + p)^{0.37} \quad (5.5)$$

4.3 micron CO_2 band $\nu_1 = 2160$ cm^{-1}, $\nu_2 = 2500$ cm^{-1}

$$\int A_\nu d\nu = 27.5 + 34 \log w + 31.5 \log (P + p) \quad (5.6)$$

2.7 micron CO_2 band $\nu_1 = 3480$ cm^{-1}, $\nu_2 = 3800$ cm^{-1}

$\int A_\nu d\nu > 50$ cm^{-1}

$$\int A_\nu d\nu = -137 + 77 \log w + 68 \log (P + p) \quad (5.7)$$

$$\int A_\nu d\nu < 50 \text{ cm}^{-1} \quad \int A_\nu d\nu = 3.15\, w^{1/2} (P + p)^{0.43} \quad (5.8)$$

6.3 micron H_2O band $\nu_1 = 1150$ cm^{-1}, $\nu_2 = 2050$ cm^{-1}

$\int A_\nu d\nu > 160$ cm^{-1}

$$\int A_\nu d\nu = 304 + 218 \log w + 157 \log (P + p) \quad (5.9)$$

$$\int A_\nu d\nu < 160 \text{ cm}^{-1} \quad \int A_\nu d\nu = 356\, w^{1/2} (P + p)^{0.30} \quad (5.10)$$

3.2 micron H_2O band $\nu_1 = 2800$ cm^{-1}, $\nu_2 = 3340$ cm^{-1}

$$\int A_\nu d\nu = 40.2\, w^{1/2} (P + p)^{0.30} \quad (5.11)$$

2.7 micron H_2O band $\nu_1 = 3340$ cm^{-1}, $\nu_2 = 4400$ cm^{-1}

$\int A_\nu d\nu > 200$ cm^{-1}

$$\int A_\nu d\nu = 337 + 246 \log w + 150 \log (P + p) \quad (5.12)$$

$$\int A_\nu d\nu < 200\ \text{cm}^{-1} \qquad \int A_\nu d\nu = 316\ w^{1/2}\ (P + p)^{0.32} \tag{5.13}$$

1.87 micron H_2O band $\nu_1 = 4800\ \text{cm}^{-1}$, $\nu_2 = 5900\ \text{cm}^{-1}$

$\int A_\nu d\nu > 275\ \text{cm}^{-1}$

$$\int A_\nu d\nu = 127 + 232 \log w + 144 \log (P + p) \tag{5.14}$$

$$\int A_\nu d\nu < 275\ \text{cm}^{-1} \qquad \int A_\nu d\nu = 152\ w^{1/2}\ (P + p)^{0.30} \tag{5.15}$$

1.38 micron H_2O band $\nu_1 = 6500\ \text{cm}^{-1}$, $\nu_2 = 8000\ \text{cm}^{-1}$

$\int A_\nu d\nu > 350\ \text{cm}^{-1}$

$$\int A_\nu d\nu = 202 + 460 \log w + 198 \log (P + p) \tag{5.16}$$

$$\int A_\nu d\nu < 350\ \text{cm}^{-1} \qquad \int A_\nu d\nu = 163\ w^{1/2}\ (P + p)^{0.30} \tag{5.17}$$

1.1 micron H_2O band $\nu_1 = 8300\ \text{cm}^{-1}$, $\nu_2 = 9300\ \text{cm}^{-1}$

$$\int A_\nu d\nu = 31\ w^{1/2}\ (P + p)^{0.26} \tag{5.18}$$

From these formulas it is a simple matter to determine the approximate amount of absorption in a given path length and to estimate the path lengths required for a significant amount of absorption to occur within a given band. In most cases the partial pressure term, p, due to the absorbing gas can be neglected compared to the total atmospheric pressure, P. For the carbon dioxide bands the calculation is simplified by the fact that the carbon dioxide content of the atmosphere is reasonably homogeneous and constant. Based on a carbon dioxide content of the atmosphere of 0.033% by volume, the curves given in Figure 26 were calculated. The amount of CO_2, w, along the path in atmosphere-cm is calculated as $33 \times 10^{-5} \times P$ (atmospheres) $\times\ l$ (cm) where is the path length. These curves give the amount of absorption in each CO_2 band as a function of the path length for a horizontal, constant pressure path in the atmosphere. From the curves one can determine the useful range for each CO_2 absorption band in the atmosphere. For the case of water vapor the situation is considerably more complicated since the water vapor content of the atmosphere is so highly variable in time and space. The quantity w in the formulas for water vapor is the amount of precipitable water vapor along the path in cm if condensed into liquid water. If one knows the water vapor density, ρ_w, along the path in gm per cm^3, then the precipitable water vapor is this quantity divided by the density of liquid water, which is 1.0 gm per cm^3, and multiplied by the path length in cm. Explicitly as an equation, one has

$$w(\text{cm}) = \frac{\rho_w(\text{gm per cm}^3)}{1\,(\text{gm per cm}^3)} \times l\,(\text{cm}) \qquad (5.19)$$

Usually one does not know ρ_w directly but is more likely to know the relative humidity and the air temperature, or the wet-bulb and dry-bulb temperature readings. From the relative humidity, which by definition is the ratio of the actual vapor pressure to the vapor pressure at saturation, one can get the actual vapor pressure.

$$\text{r.h.} = p/p_{\text{sat.}} \qquad (5.20)$$

The value of $p_{\text{sat.}}$ for any given air temperature can be looked up in tables. The gas equation

$$\rho_w = \frac{p}{RT} M_w \qquad (5.21)$$

permits one to calculate ρ_w. The molecular weight of water, M_w, is 18.00; the universal gas constant, R, is 8.31×10^7 ergs per °K per mole; and T is the absolute temperature in °K. The vapor pressure, p, must be in millibars or dynes per cm^2.

Because water vapor may exist in the atmosphere under conditions ranging from very dry to very moist, one cannot use a single curve to represent the relationship between absorption and path length, but must give a family of curves for various vapor pressures for each absorption band. If the investigator does not know the vapor pressure but knows the relative humidity and the air temperature, then the vapor pressure can be calculated by means of the procedure suggested above. Figures 27, 28, 29, 30, and 31 show the total absorption in the water vapor bands for various path lengths at constant vapor

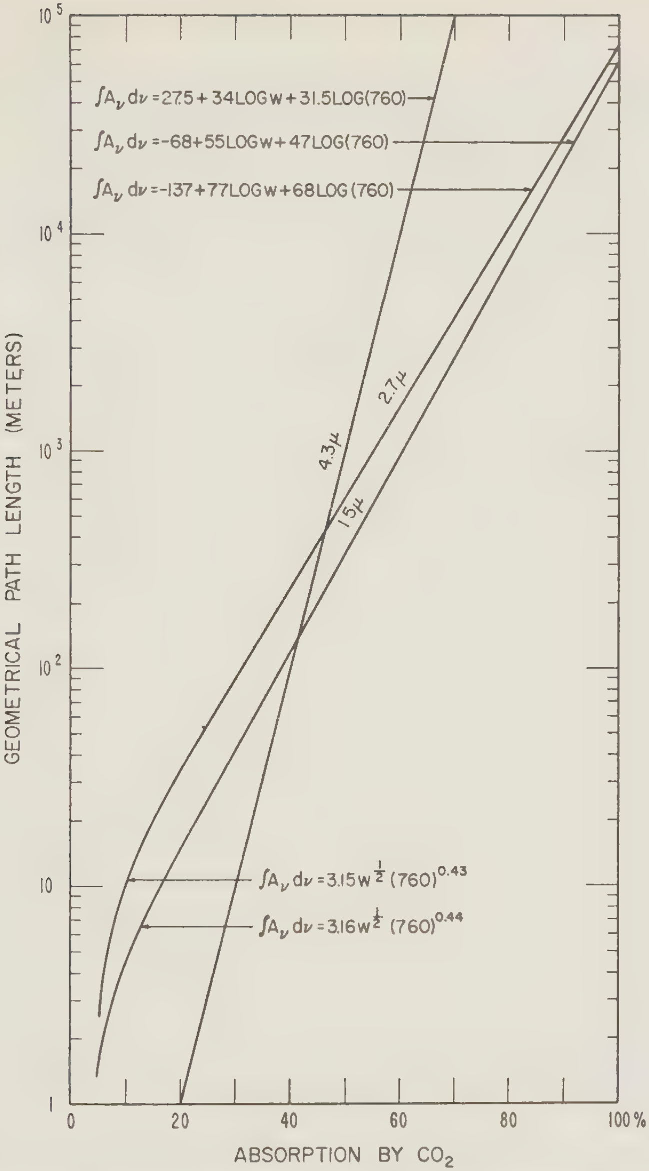


FIGURE 26.—The total absorption in the 2.7, 4.3, and 15 micron CO_2 bands as a function of the path length along a constant pressure path with CO_2 content 0.03% by volume.

pressure amounts. It should be noted how very different are the strengths of some of the water vapor bands and by this property one can select the band most useful for absorption measurements along a given path. If, either by means of filters or with the use of monochromators, spectral intervals narrower than the entire band width are selected, considerably shorter path lengths will be required to produce a substantial amount of absorption.

A light source, such as a headlight beam or spotlight, emits sufficient infrared radiation so that it can be readily detected at distances from a few meters to several hundred meters. A hygrometer of this type has been designed by Wood (1958, 1959) in which interference filters are used to select a narrow band of radiation at 2.35 microns which is unabsorbed and a narrow band at 2.70 microns which is absorbed by water vapor. An uncooled lead sulphide cell serves as detector. The light beam is chopped at a frequency of 120 cycles per second in order to provide a modulated signal for amplification. The ratio between the signal received in the 2.60 micron band and that received in the 2.35 micron band is a direct measure of the amount of water vapor in the path between the detector and the light source. According to Wood, the use of a fixed path length system appears capable of achieving an accuracy of about 5% over a density range of water vapor concentration amounting to two orders of magnitude. An earlier version of this type of instrument for water vapor measurements was developed by Foskett and Foster (1943) of the United States Weather Bureau, who used a small diffraction-grating spectrometer, rather than a filter technique, for obtaining the proper wavelength separation.

Two towers could be set up within a stand of vegetation, separated by 50 or 100 meters, and the light source could be mounted on one tower and received by the filter hygrometer mounted on the second tower. This setup would give a single path of light through the vegetation. Multiple paths of light could be achieved either by mounting several light sources on one tower and several receivers on the second tower, or by the use of mirrors which would reflect the light from a

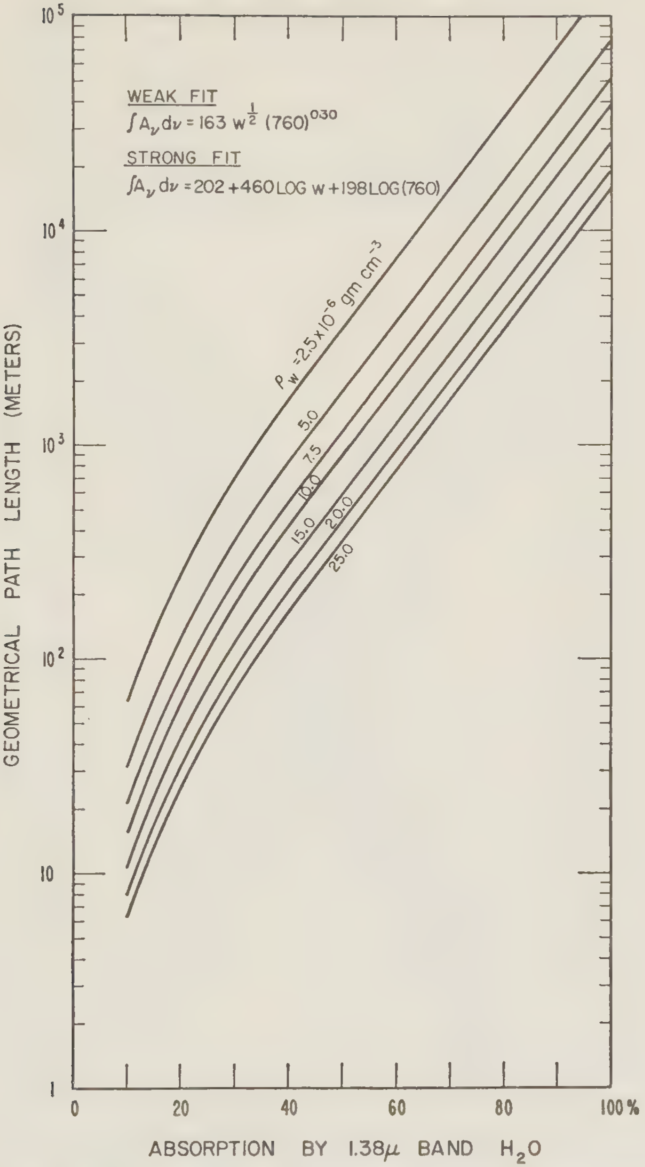


FIGURE 27.—The total absorption in the 1.38 micron H_2O band as a function of the path length along a constant pressure path with moisture density ρ_w.

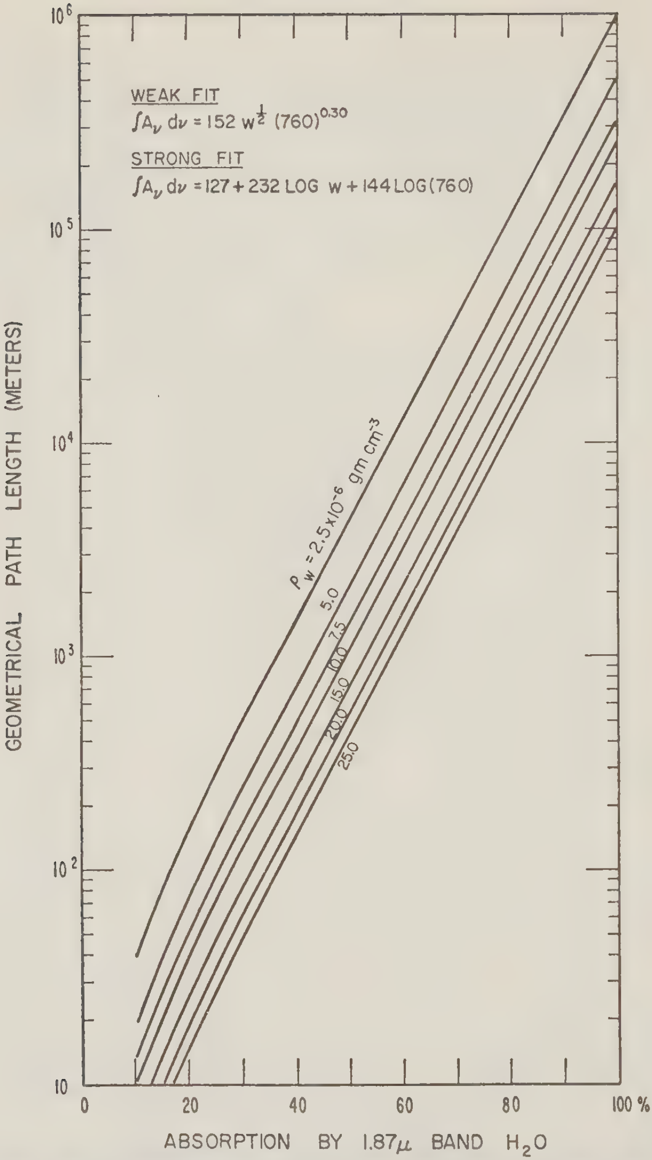


FIGURE 28.—The total absorption in the 1.87 micron H_2O band as a function of the path length along a constant pressure path with moisture density ρ_w.

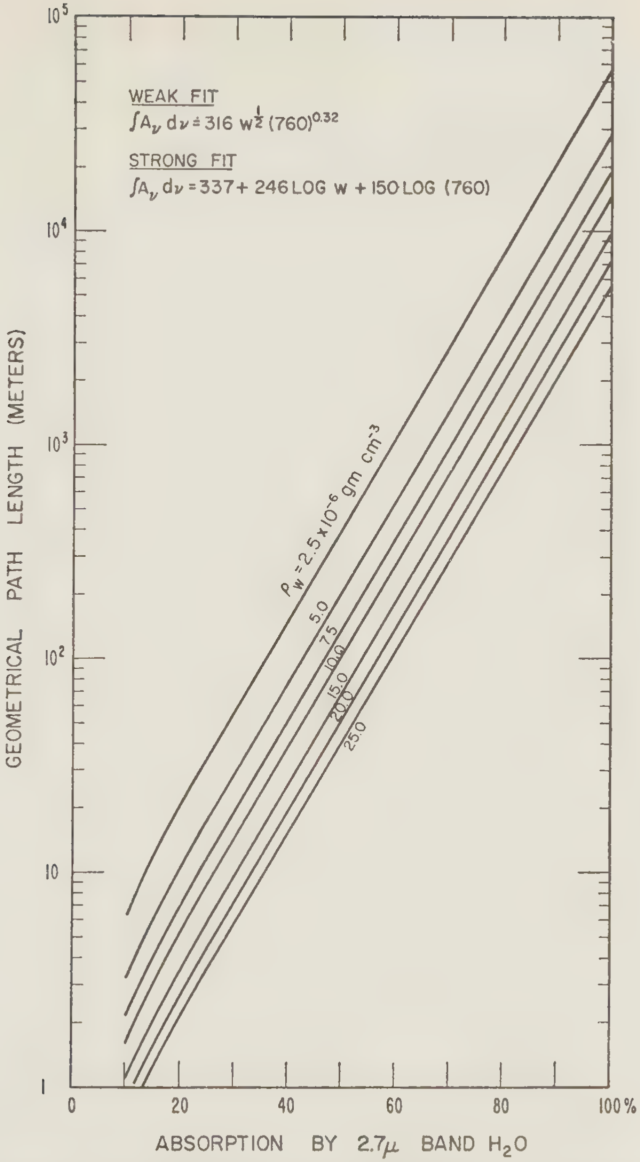


FIGURE 29.—The total absorption in the 2.7 micron H_2O band as a function of the path length along a constant pressure path with moisture density ρ_w.

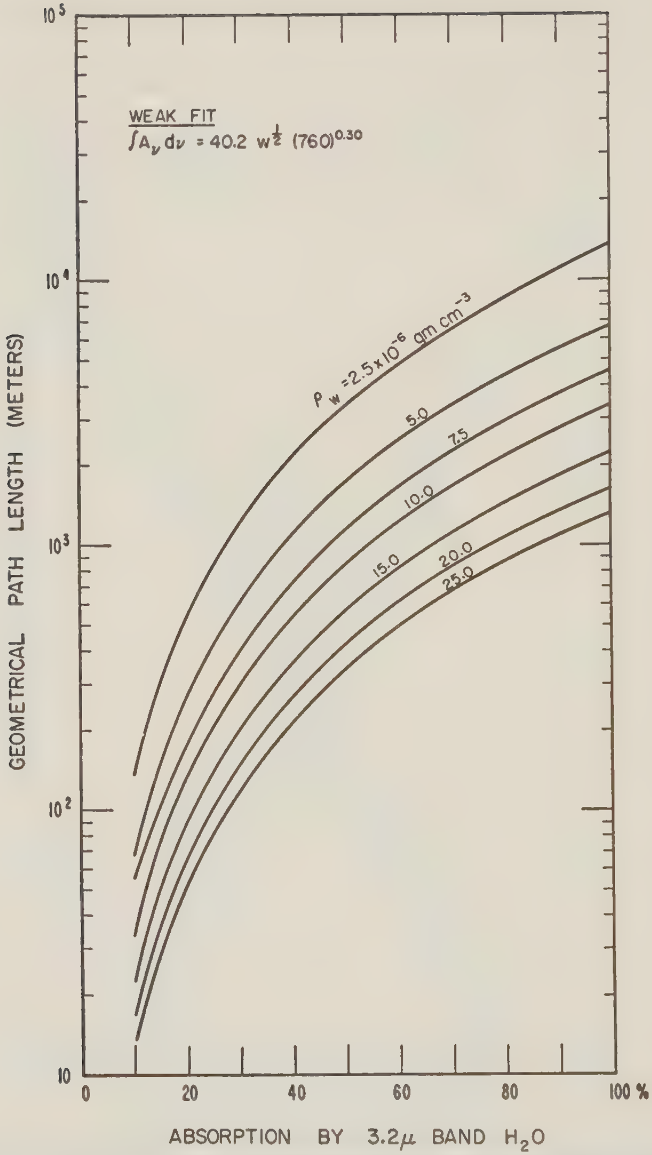


FIGURE 30.—The total absorption in the 3.2 micron H_2O band as a function of the path length along a constant pressure path with moisture density ρ_w.

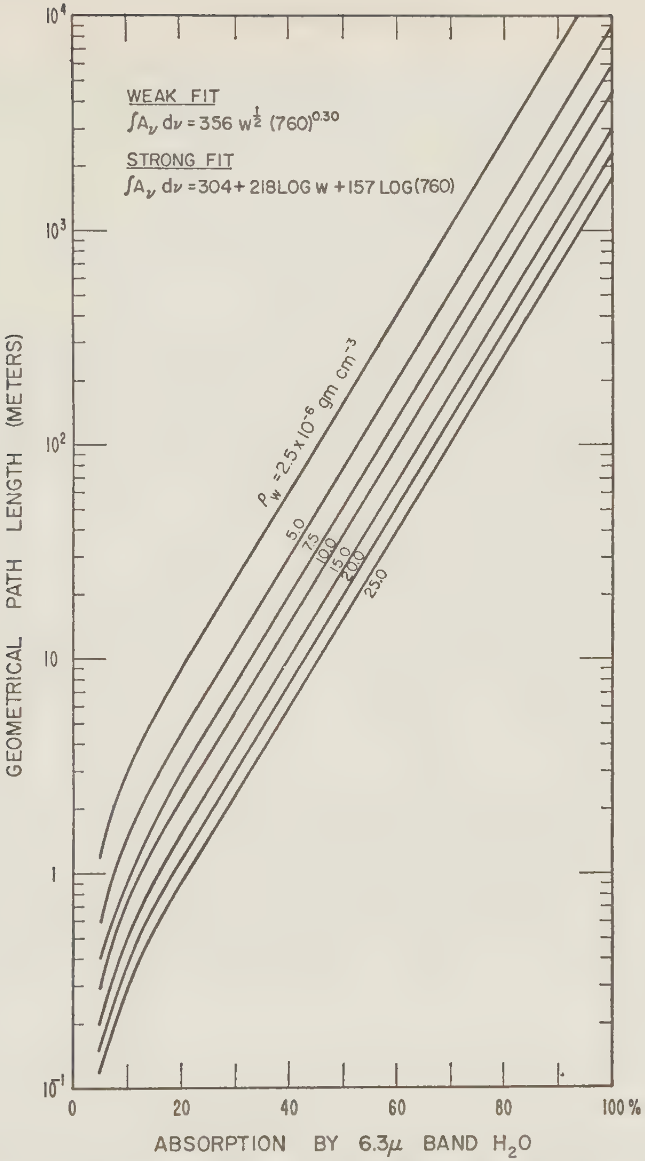


FIGURE 31.—The total absorption in the 6.3 micron H_2O band as a function of the path length along a constant pressure path with moisture density ρ_w.

single source across different heights to a single receiver on the other tower. Multiple horizontal beams at different heights are desirable since this will give one a vertical profile of the water vapor content within or above a stand of vegetation. It is important to know the vertical profile of water vapor or carbon dioxide content of the air and the wind profile, in order to be able to compute the mass transfer of these quantities to or from a stand of vegetation.

It would be important to make measurements of this type, where the total water vapor or carbon dioxide content along a horizontal path is measured within or above stands of vegetation varying from a coniferous or deciduous forest, to a grass cover, to a desert situation or above a lake surface, a bog surface, or over sand dunes or lake shore.

Infrared gas analyzers are commercially available and are commonly employed in industry for the control and analysis of plant processes. These same analyzers can be applied to the analysis of the carbon dioxide content of a crop or stand of vegetation. The gas to be analyzed must be drawn into a tube from some particular position within the crop and transported to the analyzer, and thus in this way it will only determine the CO_2 content of the air at this particular position within the crop. This technique has been employed by Monteith and Szeicz (1960) for the determination of the carbon dioxide content within a crop of sugar beets. Similar techniques have been applied by Decker (1954, 1959a,b) and by Decker and Wien (1960) for determining the CO_2 or H_2O exchange in the photosynthetic process. Decker employed a Liston-Becker model 15A infrared gas analyzer which was sensitized to either CO_2 or H_2O.

RATE OF HEAT TRANSFER WITHIN A SUBSTANCE

It takes time for heat to be transferred to the interior of an object. The velocity with which a heat wave may be propagated into an object depends upon the conductivity, the specific heat, the density, and the frequency or the wavelength of the heat wave. If heat is applied to the plane surface of a body in a periodic fashion, the rate at which heat penetrates from the surface into the interior and affects the temperature, T, at any depth, x, is given by the differential equation:

$$\frac{\partial^2 T}{\partial x^2} = \frac{\rho(S.H.)}{\kappa} \cdot \frac{\partial T}{\partial t} = \frac{1}{n} \cdot \frac{\partial T}{\partial t} \tag{5.22}$$

where ρ is the density of the material, $S.H.$ the specific heat, κ the thermal conductivity, and n the diffusivity. A particular solution to this equation can be obtained when the average temperature at the surface is given by T and periodic heating and cooling causes the surface temperature to vary by $\Delta T \sin \omega t$. Here ΔT is the amplitude of the periodic surface temperature variation. $\omega = 2\pi f$ where f is the frequency in cycles per sec of the periodic heating and t is the time. The solution to Equation 5.22 is:

$$T = T_H - \frac{\partial T}{\partial t} \cdot x + \Delta T e^{-x\sqrt{\omega/2n}} \sin \left(\omega t - x\sqrt{\frac{\omega}{2n}}\right) \tag{5.23}$$

At any depth, x, the temperature will have a maximum value when:

$$\sin \left(\omega t - x\sqrt{\frac{\omega}{2n}}\right) = 1 \tag{5.24}$$

The temperature wave will have successive maxima and minima with depth, each decreasing in amplitude. The distance between two successive maxima or minima is called the wavelength of the wave, λ. At any instant one has

$$\left[\omega t - x\sqrt{\frac{\omega}{2n}}\right] - \left[\omega t - (x + \lambda)\sqrt{\frac{\omega}{2n}}\right] = 2\pi = \lambda\sqrt{\frac{\omega}{2n}} \tag{5.25}$$

from which one gets

$$n = \frac{\omega\lambda^2}{8\pi^2} = \frac{f\lambda^2}{4\pi} = \frac{\lambda^2}{4\pi P} = \frac{\kappa}{(S.H.)\rho} \tag{5.26}$$

Thus a measurement of the wavelength, λ, and the period of the wave gives the value of the diffusivity, n, from which one can also get the conductivity of the media, κ.

The velocity of propagation of a wave, v, is the wavelength divided by the period, hence:

$$v = \frac{\lambda}{P} = \frac{\kappa}{(S.H.)\rho} \cdot \frac{4\pi}{\lambda} = \frac{4\pi n}{\lambda} = \sqrt{\frac{4\pi n}{P}} \tag{5.27}$$

The time, t, for the temperature maxima or minima to move a distance, x, in the media is given by:

$$t = \frac{x}{v} = x\sqrt{\frac{P}{4\pi n}} \tag{5.28}$$

For the surface of the earth one can calculate the velocity of penetration of the diurnal heating due to sunlight incident on the surface. An approximate value for the diffusivity, n, of soil is 0.005 cm^2 per sec. The velocity of propagation for the diurnal wave into the soil is:

$$v = \sqrt{\frac{4\pi(0.005)}{86{,}400}} \doteq 8.53 \times 10^{-4} \text{ cm per sec} \tag{5.29}$$

From this one can calculate that it would take 32.5 hours for the wave to penetrate the soil to a depth of 1 meter.

The heat load on the surface of a tree trunk or branch penetrates into the interior at the rate of approximately 1.7 cm per hr. It would take 3 hours to penetrate 5 centimeters and 24 hours to go 40 centimeters. A small twig of 1 mm radius will respond to the exterior heat load in 3.5 minutes. A trunk or branch of radius 20 centimeters would have its warmest temperatures at the center 12 hours out of phase with the temperature maximum at the surface, i.e., in the middle of the night. The heat wave entering wood will be damped in amplitude as it travels in depth until at a depth of about 10 centimeters it is less than one-fifth of its value at the surface. Hence, if there was a 5°C change in temperature at the surface, there will be only a 1°C change at 10 centimeters depth. It is clear that even the sudden short-period changes in heat load produced by sun and clouds in the daytime will influence the entire volume of a twig or small branch, but the larger branches and trunks of trees will respond primarily to the diurnal and annual changes in heat load and will be relatively unaffected by short-term pulsations of the order of minutes to an hour. A small twig also will be quickly influenced by convective cooling and heating, an effect that may be stronger than the solar heating, particularly if there is a wind. Pine needles should seldom depart very much from the ambient air temperature due to convective cooling or heating even in still air. Reference to Equation 4.22 (p. 107) will show the strong effect of convection on cylinders when the diameter of the cylinder becomes very small as it does for pine needles. The broader surface of deciduous leaves acts as an efficient collector of sunlight and in turn reduces the strong effect of convection; therefore such leaves may become considerably warmer than the ambient air. In addition, strong transpiration of such leaves may cool them below ambient air temperature when the heat load is weak. It is interesting to note that the trunk of birch trees is white and reflective in the visible, but the smaller twigs on the tree are brown, where the surface characteristics are relatively meaningless by virtue of the small diameter and convective influence. Perhaps this is not the reason for the lack of white bark on the twigs, but in any case it is important to note that coloration would not have any significant effect here.

THE MEASUREMENT OF LEAF TEMPERATURES

Although it is possible to determine the interior temperature of leaves by means of thermocouples inserted into the leaf, this method is slow, time consuming, and subject to disturbance of the leaf. A new method will be described here, which is fast and permits one to determine the surface temperature of the leaf without making contact with the leaf itself. This method makes use of the fact that a leaf radiates energy very closely to that radiated by a black body. The emissivity of deciduous leaves in the infrared at 10 microns wavelength is approximately 0.95, and the leaves radiate infrared energy according to the law σT^4. By measuring the quantity of energy radiated one can then solve for the temperature of the radiating surface.

For this type of measurement, an infrared radiometer is required. One such instrument is the Stoll-Hardy radiometer described in Chapter 3. This instrument has a field of view of 20° and an aperture of about 15 mm. If it is held very close to the leaf surface, it can determine the temperature of a circular area 15 mm or more in diameter. This permits easy measurement of most deciduous leaf temperatures. A measurement of the upper and lower surfaces of the leaf can be taken in quick succession. The accuracy of these measurements should be about 0.3°C. The instrument can also be used to measure the temperature of leaves or needles which do not fill the field of view if a mass of vegetation is available so that one leaf backs up another. Similarly for small plants forming a ground cover, it is possible to obtain a temperature characteristic of the entire mass. When pointed at a single tree, a forest, or a thicket of vegetation, the radiometer will give the mean temperature of the leaves providing it does not see through to sky or ground on the other side.

Measurements made with this type of instrument only determine the surface temperature of the leaf since the infrared radiation originates from near the surface. It might be argued that the temperature within the leaf is more important from the standpoint of cellular activity. However, the interior temperature of the leaf will be between the temperatures of the two surfaces. For very thin leaves the temperatures of the two surfaces may be very nearly the same, and this will also represent the interior temperature. For thick leaves the upper and lower surfaces of the leaf may differ by several degrees in temperature, and in this instance the interior temperatures will fall on a gradient between the two. Even in this situation it is better to know the surface temperatures since they will bracket the interior temperature. The instrument actually measures for a leaf the quantity which one needs to know for energy considerations. This quantity is the energy radiated as infrared radiation by the surface. If the energy incident on the leaf can be determined and the energy exchange by convection estimated from the difference between leaf temperature and air temperature, then the energy lost by transpiration can be calculated. An example is given below.

The temperatures and radiation emitted by the leaves of many different plants during summer months have been determined by the author using the radiometer. Leaves which are strongly sunlit are always considerably warmer than the air, and during the daytime the shade leaves are always cooler than the air. At night, leaves generally are cooler than the air by one or two degress centigrade. Measurements made on Quercus macrocarpa gave sunlit leaf temperatures commonly from 10°C to 12°C above the ambient air temperature with extremes as high as 17°C in still air. Other species, such as Ulmus glabra, Ulmus pumila, Acer plantanoides, and Populus acuminata, exhibited similar temperatures when exposed to the full sunlight in still air between 1000 and 1400 hours during July. Populus tremuloides consistently showed temperatures slightly lower, running normally 8°C above ambient air temperature and with a maximum difference of 11°C.

The direct sunlight and skylight, visible and near infrared, received on a horizontal surface as measured with a Robitzsch type of pyrheliometer was 1.60 cal per cm^2 per min. In addition the radiation from the sky added approximately 0.40 cal per cm^2 per min as estimated from observations with the radiometer pointed at the sky. The total heat load on the upper surface of the leaf was approximately 1.60 cal per cm^2 per min. The grass-covered ground reflected a small amount of sunlight and skylight upward, estimated to be 0.10 cal per cm^2 per min. The warm grass-covered ground itself (40°C) radiated upward 0.80 cal per cm^2 per min. The total heat load on the lower surface of the leaf was approximately 0.90 cal per cm^2 per min. Each element of leaf comprised of 1 cm^2 of upward facing area and 1 cm^2 of downward facing area received a total of approximately 2.50 cal per cm^2 per min. These 2 cm^2 of leaf surface area at a temperature of 46°C when the air temperature was 34°C radiated 1.68 cal per min. From our discussion in Chapter 4 concerning convection, one can make an estimate of the maximum amount of energy transferred by natural convection. From Equation 4.4 (p.

98), one finds for a leaf which is 12°C warmer than the air, treated as a horizontal plate, that the upper surface will lose approximately 0.012 cal per cm^2 per min by convection and the lower surface will lose 0.006 cal per cm^2 per min, giving a total loss of 0.018 cal per cm^2 per min from the two surfaces. It is difficult to see how these numbers can be very much greater. The energy balance equation may then be written as follows in terms of the notation of Chapter 1.

$$S_u(l - t - r) + S_l + R_u + R_l + LE + C_u + C_l = 0$$

$$1.60\ (1 - 0.20) + 0.10 + 0.40 - 0.84 + 0.80 - 0.84 + LE - 0.012 - 0.006 = 0$$

$$2.58 - 1.70 + LE = 0$$

$$LE = 0.88 \text{ cal per 2 cm}^2 \text{ per min}$$

Each surface of the leaf must on the average transpire away energy through moisture at the rate of 0.44 cal per cm^2 per min. Taking the heat of evaporation to be 600 cal per gram, this would require the loss of 0.0015 grams per 2 cm^2 per min or 0.090 grams per cm^2 per hr. If this transpiration rate appears to be high it must be remembered that the leaves under discussion are the warmest leaves on the tree. The average transpiration rate for the tree as a whole will be very much less since the average heat load on all the leaves will be much lower than the situation represented here. Nevertheless, it is difficult to see where the amount of energy transfer for those leaves exposed to the full sunlight in still air will be much different from the values used here. It would be unreasonable to change any one of the numbers by as much as 30% of its value at the very most, and this would scarcely invalidate the estimate of the transpiration rate given here.

The measurements of leaf temperature by means of the infrared radiometer show that the leaf exposed to full sunlight undergoes an immediate and precipitous drop in temperature when the very slightest breeze, less than 2 mph, occurs. For a strong breeze the sunlit leaf temperature drops nearly to air temperature. A typical example is the case of sunlit Populus acuminata leaves when the temperature of the leaf was 10°C above the air temperature with no breeze and, with a breeze of about 3 mph, the temperature dropped to about 4°C above air temperature. The air temperature in this instance was 33.5°C. A similar behavior is observed when the sun is obscured by a dense cloud. As an example, during an observation of Quercus macrocarpa, the temperature of the sunlit leaves in still air was 49°C and the air temperature was 28°C. A dense cumulus cloud came across the sun and the leaf temperature dropped to 30.5°C in about 30 seconds. Comparisons were also made between the temperature of the upper and the lower surfaces of leaves. In many instances for most species, the temperature was the same within ±0.2°C for the upper and the lower surfaces. However, for Quercus macrocarpa many measurements under strong solar insolation showed the temperature of the upper sunlit surface to be as much as 2°C higher than the temperature of the lower surface. This difference can be accounted for by the stomatal transpiration taking place primarily from the lower surface of the leaf. For Elmus americana the upper sunlit surface was occasionally 1°C warmer than the lower. These observations are given here primarily to illustrate the use of the infrared radiometer for obtaining leaf temperatures as well as the radiant temperature of the environment.

Many more measurements of this type need to be made. Interesting observations of dew and frost formation on leaves could be made with this technique.

ARTIFICIAL GROWTH CHAMBERS

The use of artificial growth chambers for the study of plant response is very important, but it requires that the problem under study be carefully defined. There are essentially two choices with respect to the use of growth chambers. On the one hand they may be employed to study the behavior of plants growing in a very artificial environment which only bares certain general relationships with respect to natural environment, and on the other hand one may attempt to duplicate in the growth chamber the precise conditions occurring in nature. The first situation is relatively easy to attain, and the latter is extremely difficult to attain. Unfortunately the difference between these two extremes has not always been recognized. Sometimes the research worker has expected to achieve a fairly close simulation of nature while actually in fact he has been wide of the mark. Perhaps the most common error is to believe that by achieving an air temperature identical with the air temperature of the natural environment one is approximating the natural situation. Even if the investigator were able to produce illumination similar to sunlight, the walls of the chamber affecting the thermal infrared radiation fluxes would by themselves strongly alter the environmental conditions. The plants would not be exchanging heat to the same sources and sinks of radiation as occur in the natural surroundings—the sky, the clouds, the ground, and nearby vegetation. My objection to investigators not simulating the natural environment is not so much that they should do so, but rather they should clearly recognize the departure from the natural situation and thoroughly understand with what it is they are truly dealing. The instrumentation of growth chambers seldom permits a continual monitoring of the total heat load on some reference surface in addition to certain standard photometric measurements which are usually made. Seldom is the growth chamber instrumented for determining the flow pattern of the air inside or the strength of the circulation. Often one or two variables will be measured, the response of the plant correlated with these, and the other variables ignored.

To correctly and thoroughly simulate the natural environment is an extremely difficult task indeed. A quasi approach to the problem is made by Went (1957), wherein the growth chamber, in this case the climatron, permits sunlight to come through the plastic roof and is designed to allow strong ventilation through the louvred sides. The sunlight received at the earth's surface is the most difficult quantity to simulate of all the environmental factors because of its nonlinear and distorted spectral distribution and its diurnal, monthly, and seasonal changes. It is not sufficient to duplicate the total lumens received on a surface, but it is necessary to obtain the correct wavelength distribution for the lumens received. A great deal of effort will be required before good solar sources for laboratory use are designed; however, two recent attempts have been made by Hirt, Schmitt, Searle, and Sullivan (1960) and by Elliott, Hysell, and Meikleham (1960). The latter paper deals with the full spectrum from 300 Å to 13,000 Å and shows how the water vapor absorption bands in the near infrared can be simulated with a laboratory sun and the proper type of filters. The paper by Hirt *et al.* deals with the ultraviolet and visible region from 2700 Å to 4000 Å. The xenon arc combined with a Corning 7740 filter was found to be the best approximation to sunlight in the ultraviolet and blue end of the spectrum. The work of Elliott *et al.* essentially confirms the earlier work by Davis and Gibson (1931) in which filters were used for converting tungsten light sources at various temperatures from 2300°K to 4000°K to several different solar spectra, summer noon, winter noon, mean sun, sun outside of earth's atmosphere, etc. Elliott *et al.* constructed a laboratory sun using a General Electric 750 T 12/34 projector lamp at 2840°K placed near the focal point of an 8-inch parabolic mirror with water-cooled liquid filters of the Davis and Gibson type No. 32 in combination with a Jena BG17 glass filter.

From these papers it is evident that scientists testing solar cells and making photochemical decomposition studies are particularly concerned with the spectral distribution in artificial sunlight and feel that they must have an accurate spectral match between artificial and natural sunlight for the spectral regions of importance to the materials used. Similarly the biologist must have great concern with the spectral absorption characteristics of the plant and animal materials with which he experiments, and in particular he must be careful to match his source to the natural sunlight within these spectral regions. As an example, chlorophyll "a" has sharp absorption peaks at 4300 Å and at 6700 Å and comparatively low absorption elsewhere throughout the visible. The other varieties of chlorophyll have similar characteristics but with the absorptions shifted in wavelength. Bacteriochlorophyll has absorption peaks at 3750 Å in the ultraviolet and at 7550 Å in the infrared. All organic materials have numerous absorption bands throughout the infrared spectrum from the visible to 15 microns. The infrared absorption bands of chlorophyll, carotin, and xanthophyll were first reported on by Coblentz and Stair (1933) and more recently for chlorophyll by Weigl and Livingston (1953). All vegetation will show absorption bands due to the strong liquid water within the cells. The water bands are well known and are essentially at the wavelength positions shown earlier for the water vapor in the earth's atmosphere.

Concerning the importance of the infrared absorption bands which lie at wavelengths beyond about 1.5 microns, for organic materials it is doubtful if the energy they absorb is directly responsible for any chemical reactions. As far as is known the energy absorbed by the infrared absorption bands of the material is primarily converted to heat and thereby increases the temperature of the substance. However, this does not mean that one should neglect the importance of these absorptions, for the very fact that energy is effectively absorbed in this way and converted to heat means that the chemical rate processes which are temperature dependent will be influenced by the energy absorbed. Furthermore, the infrared absorption bands at wavelengths greater than 4.0 microns fall in the spectral region where the thermal radiation from the sky, clouds, ground, and vegetation has its main energy distribution. The depth to which long wavelength radiation from 4.0 to 30.0 microns penetrates an organic substance depends upon the strength and location of the absorption bands. For most organic materials the infrared wavelengths will be absorbed and reradiated from very near the surface of the material. Much more research is required for the infrared spectral region with regard to the interaction of radiation and chemical processes within living cells.

The next most difficult feature to simulate within artificial growth chambers are the streams of infrared thermal radiation emitted by the atmosphere. It is doubtful that one needs to duplicate this energy in terms of its exact spectral characteristics, but at least one should produce equal amounts of total energy flux at infrared wavelengths. The nighttime sky, if clear and dry, will represent a strong sink of radiation to all organisms exposed to it. The environmental chamber should be capable of simulating this cold sink for radiation. This can be done by having a hemispherical dome with cooling coils imbedded in it so that the coldest temperature can be produced at the zenith with a warming trend toward the horizon. Sky temperatures as low as —50°C will need to be simulated. At the same time the air within the chamber must be maintained at the mean air temperature of the simulated environment. In the final analysis, if a plant is being investigated within the chamber and true environmental conditions are to be duplicated, then it is important with regard to the infrared radiation to obtain the correct total infrared flux on the plant. It will also be necessary to assure essentially the same convective exchange of energy as will be occurring in nature. This will require care with respect to the air temperature and the circulation of air within the chamber. The same type of instrumentation is required within artificial growth chambers for monitoring the energy fluxes as has been described in the earlier chapters for measuring these fluxes within the natural environment.

THE SPECTRAL PROPERTIES OF VEGETATION

The spectral properties of vegetation with regard to reflectivity, transmissivity, and absorptivity are widely variable, both with respect to wavelength and to species. The spectral characteristics of a leaf determine the net energy absorbed and therefore play an important role in the heat balance of the leaf. The spectral characteristics of the energy incident upon a leaf, whether it be sunlight, atmospheric infrared radiation, or infrared radiation from the ground, will determine the relative importance of the various spectral regions, e.g., ultraviolet, visible, or infrared.

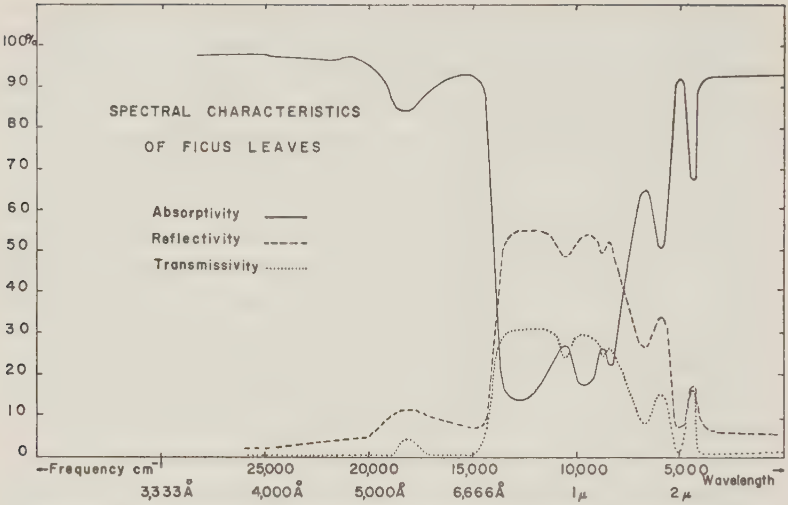


FIGURE 32.—The spectral characteristics, absorptivity, reflectivity, and transmissivity of Ficus leaves as a function of the frequency and wavelength.

A typical example of the spectral characteristics of leaves is shown for Ficus in Figure 32. It is preferred to use as an example a specific plant rather than attempting to give the average situation for many different plants in widely varying situations. The reflectivity and absorptivity values for the visible portion of the spectrum were taken from Moss and Loomis (1952). The transmissivity in the visible was taken as the difference of these from 100%. In the near infrared from 0.8μ to 2.5μ the work of Dinger (1941) was used, and for longer wavelengths the work of Gates and Tantraporn (1952). The Dinger work gave all three values—reflectivity, transmissivity, and absorptivity. The characteristic brightness of vegetation in the photographic and near infrared from 7150 Å to about 1.5μ is readily seen in the reflectivity curve of Figure 32. However, it should also be noted, as reported by Gates and Tantraporn (1952), that the brightness does not continue into the longer wavelength infrared and, indeed, at wavelengths greater than about 2.0μ, plants are almost "black." The transmissivity and reflectivity have very similar spectral functions while the absorptivity values are the conjugate of these. The absorptivity curve is a complex one as a function of the frequency or the wavelength, having a distinct minimum in the near infrared where vegetation reflects strongly.

In order to determine the heat balance for leaves in sunlight the average absorptivity of the leaf to incident solar radiation must be determined. This can be done by multiplying the incident solar energy curve as a function of the frequency or the wavelength by the spectral absorptivity values. Figures 33 and 34 show the incident solar radiation for two extreme conditions at the earth's surface and the absorbed solar energy which results when the upper curve is multiplied by the absorptivity of the leaf. Here the energy is plotted against the frequency of the radiation in wave numbers (cm^{-1}), and therefore any area under the curve represents

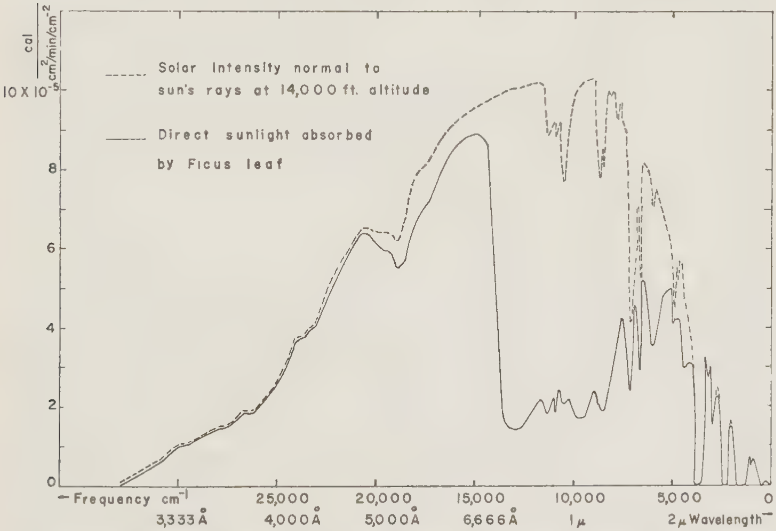


FIGURE 33.—Solar intensity at the earth's surface at 14,000 feet altitude normal to the sun's rays for the sun in the zenith (air mass 1.0). The atmosphere is very clear and very dry (1 mm of precipitable water vapor). The total energy under this curve is 1.74 cal per cm² per min. The solar energy absorbed by a Ficus leaf normal to the direct rays from the sun is shown in the lower curve. The average absorptivity to the solar energy is 63.3%.

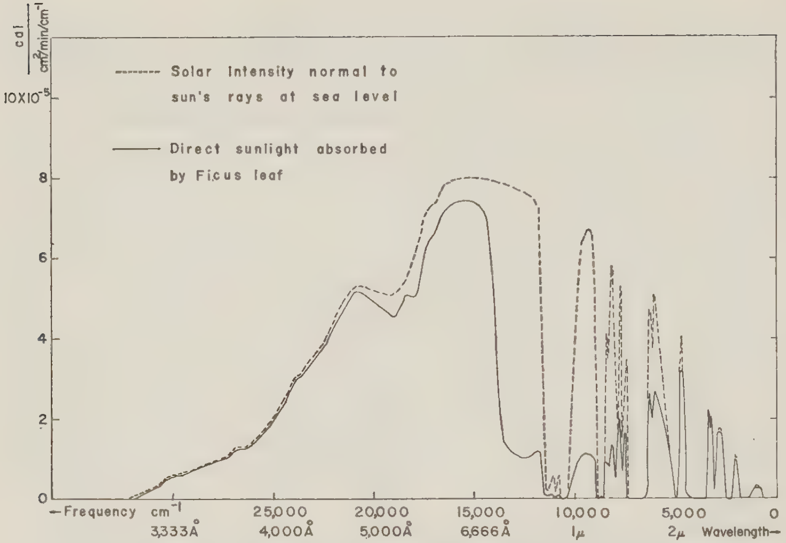


FIGURE 34.—Solar intensity at the earth's surface at sea level normal to the sun's rays for the sun in the zenith (air mass 1.0). The atmosphere is hazy and very moist (25 mm of precipitable water vapor). The total energy under this curve is 1.03 cal per cm² per min. The solar energy absorbed by a Ficus leaf normal to the direct rays from the sun is shown in the lower curve. The average absorptivity to the solar energy is 71.2%.

energy according to the well-known quantum law that $E = \hbar\nu$, where E is the energy, $\hbar$ is Planck's constant, and ν is the frequency of the radiation. The upper curve of Figure 33 represents the incident energy normal to the solar beam when the sun is in the zenith (air mass is unity) and the atmosphere is very clear and dry (1 mm of precipitable water vapor), a condition which occurs for high mountain regions. The lower curve represents the solar energy absorbed by a Ficus leaf. By taking the ratio of the total area under the absorbed energy curve to the total area under the solar energy curve the average absorptivity can be determined. For the situation shown in Figure 33 the average absorptivity for the Ficus leaf normal to the sun's rays is 63.3%. The upper curve in Figure 34 represents the incident energy normal to the solar beam when the sun is in the zenith (air mass is unity) and the atmosphere is hazy and very moist (25 mm precipitable water vapor), a condition which often occurs near sea level. The lower curve represents the solar energy absorbed by a Ficus leaf. Computing the average absorptivity in the same manner as before one gets 71.2%. Hence it is seen that the average absorptivity to sunlight differs considerably with the spectral characteristics of the sunlight, a property which will vary with the time of day, condition of the sky, season, and altitude. One definitely cannot speak of a unique absorptivity for a leaf.

The absorptivity of the Ficus leaf used in this illustration is approximately 97% at wavelengths shorter than 5000 Å, is approximately 20% from 7400 Å to 12,500 Å (1.25μ), and is 93% at wavelengths greater than 25,000 Å (2.5μ). It will be noticed by comparing Figure 32 with Figures 33 and 34 that the strong minimum in the absorptivity curve, between 7400 Å and 12,500 Å, which corresponds to maximum reflectivity and transmissivity, is uniquely located in the region of maximum incident solar energy. The absorptivity is high in precisely those regions of the visible spectrum where chlorophyll has strong absorption, namely the region from 4000 Å to 5000 Å and the region 6200 Å to 7000 Å (see Hollaender, 1956).

The absorptivity drops abruptly to the long wavelength side of the red absorption band in chlorophyll. Despite the fact that the absorptivity of a leaf is high in the regions of the spectrum where photochemical stimulation for photosynthesis is required, from the standpoint of the heat load and the resulting temperature, it is the average absorptivity to the total incident solar energy which is important. This average absorptivity for Ficus varies between 63% and 71% depending upon the sun and sky conditions. It is also a fact that laboratory experiments involving artificial sources of light will result in different total absorption than for the same plant in natural sunlight. This is true even if the total incident energy may be matched to that in natural sunlight. If the spectral distribution is not the same, the average absorptivity for the leaf will be different.

The absorptivity of leaves is also high in the infrared at wavelengths greater than 2.5μ. However, the energy from the sun is relatively weak at these longer wavelengths and does not of itself add substantially to the heat load. The high absorptivity at the longer wavelengths does mean that the leaves absorb efficiently the thermal infrared radiation coming from the sky, from the ground, and from other leaves, plants, and objects nearby. The absorption of this energy plays a substantial role in the heat balance of the plant, particularly at night but also to some extent during the day.

The excellent work of Billings and Morris (1951) on the reflectivity of leaves from different ecological groups to visible and near infrared radiation should be referred to. For the visible region of the spectrum they found the greatest reflectivity for desert species, with subalpine species next, followed by species from the more mesic sites. The reflectivities of leaves given by Shull (1929) and Clark (1949) show for the visible and photographic infrared that the lower surface of the leaf reflects considerably more than the upper surface, presumably because of the lack of palisade cells on the lower side. Gates and Tantraporn (1952) showed that the reverse was true for the long-wave infrared with the upper surfaces having higher reflectivity than the lower surfaces. These authors discussed the mechanism by which visible light is reflected from a leaf by internal reflection and scattering within the cells of the spongy parenchyma and how the longer wavelength infrared radiation is reflected primarily at the outer epidermal surface.

Other references concerning the spectral and optical properties of plant leaves are as follows: Kleshnin and Shulgin (1959), Shulgin, Khasanov, and Kleshnin (1960), and Tageeva, Brandt, and Derevyanko (1960).

References

ÅNGSTROM, A. "Ueber die Anwendung der elektrischen Kompensationsmethode zur Bestimmung der nächtlichen Ausstrahlung." Nova Acta Soc. Sci. Uppsala Ser. 4, **1,** No. 2 (1905).

———. Smithsonian Miscellaneous Collection **65,** No. 3 (1915).

———. "Note on the relation between time of sunshine and cloudiness in Stockholm, 1908–1920." Archiv. fûr Matemat. Astronom., und Physik **17,** No. 15 (1922).

———. "Actinometric Measurements." *Compendium of Meteorology,* Amer. Meteor. Soc., Boston (1951), pp. 50–57.

ÅNGSTROM, A., and ASKLOF, S. Ann., Stockholm, **2,** 253 (1920).

BAUMGARTNER, A. "Die Strahlungsbilanz in einer Fichtendickung." Forstwissenschaftl. Centralblatt **71,** 337–49 (1956a).

———. "Untersuchunger über den Wärme-und Wasserhaushalt eines jungen Waldes." Berichte des Deutschen Wetterdienstes **5,** 4–53 (1956b).

BENER, P. "Investigation on the spectral intensity of ultraviolet sky and sun and sky radiation under different conditions of cloudless weather at 1590 meters above sea level." Physikalisch-Meteorologisches Observatorium, Davos, Switzerland. Technical Summary Report No. 1, Contract AF 61(052)-54 (1960).

BERRY, F. A., BOLLAY, E., and BEERS, N. R. *Handbook of Meteorology,* McGraw-Hill Book Co., New York (1945).

BILLINGS, W. D., and MORRIS, R. J. "Reflection of visible and infrared radiation from leaves of different ecological groups." Amer. J. Bot. **38,** 327–31 (1951).

BLACK, J. N. "The distribution of solar radiation over the earth's surface." Archiv. Met., Geoph. Biokl. Ser. B, **7,** 165 (1956).

BLACKWELL, M. J. "On the development of an improved Robitzsch-type actinometer." Met. Res. Comm., Air Ministry, London M.R.P., No. 791 (1953).

BRAY, J. R. "An analysis of the possible recent change in atmospheric carbon dioxide concentration." Tellus **11,** 220–30 (1959).

BROOKS, F. A. *An Introduction to Physical Microclimatology,* University of California, Davis (1959), p. 264.

BRUNT, D. "Notes on radiation in the atmosphere." Quart. J. Roy. Meteor. Soc. **58,** 389–418 (1932).

BUDYKO, M. T. *Atlas of the Heat Balance,* Leningrad (1955).

———. *The Heat Balance of the Earth's Surface,* Gidrometeorologicheskoe izdatel'stvo (Leningrad). Available as a translation, PB131692, U.S. Department of Commerce, Office of Technical Services, Washington 25, D.C. (1956).

BUDYKO, M. T., BERLIAND, T. G., and ZUBENOK, L. I. "Heat balance of the earth's surface" (in Russian). Izv. Akad. Nauk. U.S.S.R. Ser. B, **3,** 17 (1954).

BURDECKI, F. "Remarks on the distribution of solar radiation over the surface of the earth." Archiv. Met., Geoph. Biokl. Ser. B, **8,** 326–35 (1956).

CALLENDAR, G. S. "The artificial production of carbon dioxide and its influence on temperature." Quart. J. Roy. Meteor. Soc. **64,** 223–40 (1938).

———. "On the amount of carbon dioxide in the atmosphere." Tellus **10,** 243–48 (1958).

CLARK, W. *Photography by Infrared* (2d ed.), Wiley, New York (1949), p. 255.

Coblentz, W. W., and Stair, R. "Infrared absorption spectra of some plant pigments." J. Research **11,** 703–11 (1933).

Courvoisier, P. "Über den Eichfaktor des Aktinographen Robitzsch." Verhandlungen der Schweizer. Naturforschenden Gesellschaft, Zurich (1946), pp. 86–87.

———. "Über einen neuen Strahlungsbilanzmesser." Verhandlungen der Schweizer. Naturforschenden Gesellschaft, Davos (1950), pp. 152–53.

Courvoisier, P., and Wierzejewski, H. "Das Kugelpyranometer Bellani." Arch. Met. Wien. **5,** 413 (1954).

Davis, R., and Gibson, K. S. "Filters for the reproduction of sunlight and daylight and the determination of clear temperature." Misc. Pub. National Bureau of Standards, No. 114 (1931).

Decker, J. P. "The effect of light intensity on photosynthetic rate in Scotch Pine." Plant Phys. **29,** 305–6 (1954).

———. "Comparative responses of carbon dioxide outburst and uptake in tobacco." Plant Phys. **34,** 100–102 (1959a).

———. "Some effects of temperature and carbon dioxide concentration on photosynthesis of Mimulus." Plant Phys. **34,** 103–6 (1959b).

Decker, J. P., and Wien, J. D. "Transpirational swiges in Tamarax and Eucalyptus as measured with an infrared gas analyzer." Plant Phys. **35,** 340–43 (1960).

Dinger, J. E. "The absorption of radiant energy in plants." Ph.D. Dissertation, Iowa State University, Ames (1941).

Dorno, C. "Papers on the relation of the atmosphere to human comfort." Monthly Weather Review **54,** No. 2, 39–43 (1926).

———. "Die Abkuhlungsgrösse in verschiedenen Klimaten nach Daurregistrierungen mittels des Davoser Frigorimeters." Meteor. Zeits. **45,** 401–21 (1928).

Dunkelman, L., and Scolnik, R. "Solar spectral irradiance and vertical atmospheric attenuation in the visible and ultraviolet." J. Opt. Soc. Amer. **49,** No. 4, 356–67 (1959).

Ehlers, J. H. "The temperature of leaves of pinus in winter." Amer. J. Bot. **2,** 32–70 (1915).

Elliott, J. F., Hysell, R. E., and Meikleham, V. "Laboratory sun." J. Opt. Soc. Amer. **50,** No. 7, 713–17 (1960).

Elsässer, W. M. Atmospheric radiation tables. Meteorological Monographs, AFCRL-TR-60-236, U.S. Department of Commerce, Office of Technical Services, Washington 25, D.C. (1960).

Fleischer, R. "Die Atmosphärische Gegenstrahlung." Medizin-Meteor. Hefte No. 13, 142–48 (1958).

Foskett, L. W., and Foster, N. B. "A spectroscopic hygrometer." Bull. Amer. Meteor. Soc. **24,** 146 (1943).

Fritschen, L. J. "Construction and calibration details of the thermal transducer type net radiometer." Bull. Amer. Meteor. Soc. **41,** 180–83 (1960).

Fritschen, L. J., and van Wijk, W. R. "Use of an economical thermal transducer as a net radiometer." Bull. Amer. Meteor. Soc. **40,** 291–94 (1959).

Fritz, S., and MacDonald, T. H. "Average solar radiation in the United States." Heating and Ventilating **46,** 61–64 (1949).

Funk, J. P. "Improved polyethylene shielded net radiometer." J. Sci. Instr. **36,** 267–70 (1959).

———. "Measured radiative flux divergence near the ground at night." Quart. J. Roy. Meteor. Soc. **86,** 382–89 (1960).

GAGGE, A. P. "The linearity criterion as applied to partitional calorimetry." Amer. J. Physiol. **116,** 656–68 (1936).

———. "Standard operative temperature, a generalized temperature scale, applicable to direct and partitional calorimetry." Amer. J. Physiol. **131,** 93–103 (1940).

GAGGE, A. P., WINSLOW, C. E. A., and HERRINGTON, L. P. "The influence of clothing on the physiological reactions of the human body to varying environmental temperatures." Amer. J. Physiol. **124,** 30–50 (1938).

GATES, D. M. "Infrared determination of precipitable water vapor in a vertical column of the earth's atmosphere." J. Meteor. **13,** No. 4, 369–75 (1956).

———. "The infrared atmospheric transmission to solar radiation." J. Opt. Soc. Amer. **50,** No. 9, 867–82 (1960).

———. "Thermal radiation studies in Yellowstone National Park in winter." Science **134,** No. 3471, 32–35 (1961).

GATES, D. M., and TANTRAPORN, W. "The reflectivity of deciduous trees and herbaceous plants in the infrared to 25 microns." Science **115,** No. 2997, 613–16 (1952).

GEIGER, R. *The Climate Near the Ground,* Harvard University Press, Cambridge, Mass. (1957).

GIER, J. T., and DUNKLE, R. V. "Total hemispherical radiometers." Trans. Amer. Inst. Elect. Engrs. **70,** 339 (1951).

GOODELL, B. C. "An inexpensive, totalizing, net radiometer." J. Geophys. Res. **67** (in press).

GRÂFE, K. "Strahlungsempfang vertikaler, ebener Flächen; Globalstrahlung von Hamburg." Berichte des Deutschen Wetterdienstes **5,** 3–15 (1956).

HALTINER, G. J., and MARTIN, F. L. *Dynamical and Physical Meteorology,* McGraw-Hill Book Co., New York (1957).

HAND, I. "Solar energy for house heating." Heating and Ventilating **44,** 81–94 (1947).

———. "Weekly mean value of daily total solar and sky radiation." U.S. Weather Bureau Tech. Paper No. 11 (1949).

———. "Insolation on clear days at the time of solstices and equinoxes for latitude 42° N." Heating and Ventilating **47,** 92–94 (1950).

———. "Distribution of solar energy over the U.S." Heating and Ventilating **50,** 73–75 (1953).

———. "Insolation on cloudless days at the time of solstices and equinoxes." Heating and Ventilating **51,** 97–100 (1954).

HARDY, J. D. "The radiation of heat from the human body." J. Clinical Investigation **13,** 593–604 (1934).

HAURWITZ, B. *Dynamic Meteorology,* McGraw-Hill Book Co., New York (1941).

———. "Insolation in relation to cloud type." J. Meteor. **5,** 110–13 (1948).

HIRT, R. C., SCHMITT, R. G., SEARLE, N. D., and SULLIVAN, A. P. "Ultraviolet spectral energy distributions of natural sunlight and accelerated test light sources." J. Opt. Soc. Amer. **50,** No. 7, 706–13 (1960).

HOFMAN, G. Ein Strahlungsbilanzmesser für forstmeteorologische Untersuchunger. Forstwissenschaftl. Centralblatt **71,** 330–37 (1952).

HOLLAENDER, A. (ed.). *Radiation Biology,* Vol. 3, McGraw-Hill Book Co., New York (1956), p. 305.

HOWARD, J. N., BIRCH, D. E., and WILLIAMS, D. "Infrared transmission of synthetic atmospheres." J. Opt. Soc. Amer. **46** (1956).

I. "Instrumentation," No. 3, pp. 186–90.

II. "Absorption by carbon dioxide," No. 4, pp. 237–41.
III. "Absorption by water vapor," No. 4, pp. 242–45.
IV. "Application of theoretical band models," No. 5, pp. 334–38.

JOHNSON, F. S. "The solar constant." J. Meteor. **11,** 431–39 (1954).

KAPLAN, L. D. "The influence of carbon dioxide variations on the atmospheric heat balance." Tellus **12,** 204–8 (1960).

KELLEY, C. F., BOND, T. E., and ITTNER, N. R. " 'Cold spots' in the sky may help cool livestock." Agricultural Eng. **38,** 726–29 and 743 (1957a).

———. "Sky temperatures in the Imperial Valley of California." Trans. Amer. Geophys. Union **38,** 308–13 (1957b).

KIMBALL, H. H., and HOBBS, H. E. "A new form of thermoelectric recording pyrheliometer." Monthly Weather Review **51,** No. 5, 239–42 (1923a).

———. "A new form of thermoelectric recording pyrheliometer." J. Opt. Soc. Amer. **7,** 707–18 (1923b).

KLEIN, W. H. "Calculation of solar radiation and the solar heat load on man." J. Meteor. **5,** 119–29 (1948).

KLESHNIN, A. F., and SHULGIN, I. A. "The optical properties of plant leaves." Bot. Sci. Sect. Transl. **125,** 108–10 (1959). Transl. from Doklady Akad. Nauk. U.S.S.R. **125,** 1158–60 (1959).

KONDRATIEV, K. Y., and NILISK, H. I. "On the question of carbon dioxide heat radiation in the atmosphere." Geofisica pura e applicata **46,** 216–30 (1960).

KRINOV, E. L. "Spectral reflectance properties of natural formations." Laboratoriia Aerometodov, Akad. Nauk. S.S.S.R., Moscow, 1947. Translated by E. Belkov for the National Research Council of Canada as T-439 (1953).

LIU, B. Y. H., and JORDAN, R. C. V. "The interrelationship and characteristic distribution of direct diffuse and total solar radiation." Solar Energy **4,** 1–19 (1960).

MCFARLANE, W. V. "Experimental approaches to the functions of tropical livestock." Arid Zone Res. XI Proc. of the Canberra Symposium UNESCO, Paris (1958), pp. 227–33.

MILLER, D. H. "The influence of pine forest on daytime temperature in the Sierra Nevada." The Geo. Rev. **46,** 209–18 (1956).

MONTEITH, J. L., and SZEICZ, G. "The carbon dioxide flux over a field of sugar beet." Quart. J. Roy. Meteor. Soc. **86,** 205–14 (1960).

MORIKOFER, W. "Meteorologische Strahlungsmessemethoden." Handb. Biol. Arbmeth. Abt. II, Teil, 4005–4245 (1939).

MOSS, R. A., and LOOMIS, W. E. "Absorption spectra of leaves." Plant Phys. **27,** 370–91 (1952).

ORENDORFF, J. H. *Application of Climatic Data to House Design,* Housing and Home Finance Agency, Supt. of Documents, Washington, D.C. (1954).

PARRY, D. A. "Factors determining the temperatures of terrestrial arthropods in sunlight." J. Exp. Biol. **28,** 445–62 (1951).

PLASS, G. N. "The influence of the 15 micron carbon dioxide band on the atmospheric infrared cooling rate." Quart. J. Roy. Meteor. Soc. **82,** 310–24 (1956a).

———. "The carbon dioxide theory of climatic change." Tellus **8,** 140–54 (1956b).

———. "Effect of carbon dioxide variations on climate." Amer. J. Physics **24,** 376–87 (1956c).

———. "Carbon dioxide and climate." Scientific American **201,** 41–47 (1959).

POHL, W. "Messungen des ultraroten Strahlungsstromes in der freien Atmosphäre." Inaugural Dissertation zur Erlangung der Doktorwürde der Hohen Naturwissenschaftlichen Fakultät der Ludwig-Maximilians-Universität zu Müchen (1955).

Ramsay, J. A., Butler, C. G., and Sang, J. H. "The humidity gradient at the surface of a transpiring leaf." J. Exp. Biol. **15,** 255–65 (1938).

Raschke, K. "Ein verfahren der transpirationsbestimmung an Blattpunkten, Blatteilen und ganzen Blattern in situ." Naturwiss. **41,** 308 (1954).

———. "A method for the micro-meteorological measurement of the energy-budget of a leaf in situ." Ph.D. Thesis, The University of Poona, India (1955).

———. "Heat transfer between the plant and the environment." Annual Rev. Plant Physiol. **11,** 111–26 (1960).

Rider, N. E. "Eddy diffusion of momentum, water vapor and heat near the ground." Phil. Trans. Roy. Soc., London, Ser. A, **246,** 481–501 (1954).

Rider, N. E., and Robinson, G. D. "A study of the transfer of heat and water vapor above a surface of short grass." Quart. J. Roy. Meteor. Soc. **77,** 375–401 (1951).

Robinson, G. D. "Note on the measurement and estimation of atmospheric radiation." Quart. J. Roy. Meteor. Soc. **76,** 37–51 (1950).

Robitzsch, M. "Ueber den Bimetallaktinographen Fuess-Robitzsch." Beitr. Geophys. **35,** 387 (1932).

Savinov, S. I. "Concerning formulas for direct and scattered radiation and dependent on cloudiness." Meteor. Vestnik., Nos. 5–6 (1933).

Schüepp, W. "L'emploi des pyranographes Robitzsch en climat tropical." Bull. Serv. Meteor. Congo. Belge, Nos. 3, 5 (1952).

Schulze, R. "Ueber ein Strahlungsmessgerät mit ultrarotdwichlässigar Windschutzhaube am Meteorologischen Observatorium Hamburg." Geofisica pura e applicata **24,** 107 (1953).

Shulgin, I. A., Khasanov, V. S., and Kleshnin, A. F. "On the reflection of light as related to leaf structure." Bot. Sci. Sect. Transl. **134,** 197–99 (1960). Transl. from Doklady Akad. Nauk. S.S.S.R. **134,** 471–74 (1960).

Shull, C. A. "A spectrophotometric study of reflection of light from leaf surfaces." Botan. Gaz. **87,** 583 (1929).

Sloan, R., Shaw, J. H., and Williams, D. "Infrared emission spectrum of the atmosphere." J. Opt. Soc. Amer. **45,** No. 6, 455–60 (1955).

Smithsonian Miscellaneous Collection **114** (1951). Smith. Meteor. Tables, 6th revised edition.

Stair, R. "Ultraviolet radiant energy from the sun observed at 11,190 feet." J. Research **49,** 227 (1952).

Stoll, A. M. "A wide-range thermistor radiometer for the measurement of skin temperature and environmental radiant temperature." Rev. Sci. Instr. **25,** 184–87 (1954).

Stoll, A. M., and Hardy, J. D. "Thermal radiation measurements in summer and winter, Alaskan climates." Trans. Amer. Geophys. Union **36,** 213–25 (1955).

Suomi, V. E. "An economical net radiometer." Tellus **10,** 160–63 (1958).

Suomi, V. E., Franssila, M., and Islitzer, N. F. "An improved net-radiation instrument." J. Meteor. **11,** 276–82 (1954).

Suomi, V. E., Staley, D. O., and Kuhn, P. M. "A direct measurement of infrared radiation divergence to 160 mb." Quart. J. Roy. Meteor. Soc. **84,** 134–41 (1958).

Tageeva, S. V., Brandt, A. B., and Derevyanko, V. G. "Changes in the optical properties of leaves in the course of the growing season." Bot. Sci. Sect. Transl.

135, 266–68 (1960). Transl. from Doklady Akad. Nauk. S.S.S.R. **135,** 1270–73 (1960).

Thilenius, R. "Die konstruktion des Davoser frigorimeter." Meteor. Zeits. **48,** 254–60 (1931).

Weber, N. A. "Isothermal conditions in tropical soil." Ecol. **40,** 153–54 (1959).

Weigl, J. W., and Livingston, R. "Infrared spectra of chlorophyll and related compounds." J. Amer. Chem. Soc. **75,** 2173–76 (1953).

Went, F. W. *The Experimental Control of Plant Growth,* Chronica Botanica Co., Waltham, Mass. (1957).

Winslow, C. E. A., Gagge, A. P., and Herrington, L. P. "Heat exchange and regulation in radiant environments above and below air temperature." Amer. J. Physiol. **131,** 79–92 (1940).

Winslow, C. E. A., Herrington, L. P., and Gagge, A. P. "The determination of radiation and convection exchanges by partitional calorimetry." Amer. J. Physiol. **116,** 669–84 (1936a).

———. "A new method of partitional calorimetry." Amer. J. Physiol. **116,** 641–45 (1936b).

———. "The relative influence of radiation and convection upon the temperature regulation of the clothed body." Amer. J. Physiol. **124,** 51–61 (1938a).

———. "The reactions of the clothed human body to variations in atmospheric humidity." Amer. J. Physiol. **124,** 692–702 (1938b).

Wood, R. C. "Improved infrared absorption spectra hygrometer." Rev. Sci. Instr. **29,** 36–42 (1958).

———. "The infrared hygrometer as a potential meteorological aid." Bull. Amer. Meteor. Soc. **40,** 280–84 (1959).

Yamamoto, G. "On the radiation chart." Sci. Rep. Tôhoku University, Ser. 5, Geophy. **4,** 9–23 (1952).

Yamamoto, G., and Sasamori, T. "Measurement of atmospheric radiation." Sci. Rep. Tôhoku University, Ser. 5, Geophy. **6,** 19–31 (1954).

Index

Absorption bands
  mean fractional absorption in, 115
  of chlorophyll, 135–36
  total absorption in, 114
Absorptivity
  of atmosphere, 51
  of leaves, 138
Albedo, 60
Ångstrom compensation pyrgeometer and pyrheliometer, 75, 85
Artificial growth chambers, 133–37
Atmospheric absorption bands, 14, 15, 16, 65, 66
Atmospheric composition, 14
Atmospheric dust
  concentration of, 53
  scattering by, 52
Atmospheric radiation
  computed from analytical expressions, 67, 68
  computed from charts, 67

Bellani pyranometer, 75, 79–80
Black body, defined, 16–17
Boundary layer, 95, 98, 102
  laminar flow, 103
  turbulent flow, 103

Carbon dioxide
  absorption bands of, 15, 16, 65, 66
  atmospheric absorption of, 51
  climatic change, 19, 20
  climatic change due to, 19, 20
  concentration, 14, 117
  measurement of, 113–26
  radiation from, 18, 65
  total band absorption formulae, 116, 119
Conduction, 3, 4, 98
  in soil, 12
Convection, 3, 4, 6, 94–112
  coefficient defined, 97
  coefficient values
    forced convection, 105, 107
    free or natural convection, 101
  forced, 94, 102–12
    from a branch or cylinder, 100, 101, 107
    from a leaf, 102, 104–6, 131, 132
    in world climates, 21–27
    natural, 94, 99–102
Courvoisier net radiometer, 75, 83

Eddy transfer
  of heat, 110, 111
  of moisture, 110, 111
  of momentum, 110
Emissivity, 4, 64
Energy flow
  animal responses to, 41–46
  at earth's surface, 5, 10, 11
  in localized climates, 28–41
  in world climates, 21–27
  to and from man, 44–45
Energy flow equation, 3, 4
Energy regimes of the world, 21
Eppley pyrheliometer, 75–76
Evaporation, 3, 4, 6
  in world climates, 21–27

Filters, to simulate sunlight, 135
Franssila-Suomi net radiometer, 75, 82
Frigorimeter, 75, 89–90
Fritschen and van Wijk net radiometer, 75, 93
Funk net radiometer, 75, 92–93

Gas equation, 118
Gier and Dunkle net radiometer, 75, 82–83
Grashof number
  defined, 99
  in forced convection, 103
  in free or natural convection, 99, 100

Heat load
  on an animal, 13, 14
  on a human, 33, 34, 35, 37
  on a leaf, 28, 31, 32, 131, 132
Heat transfer rate
  in forced convection, 106, 107
  in free or natural convection, 102
  in solids, 126–29

Hofman heat-compensated net radiometer, 75, 84–85

Infrared gas analyzer, 126
Infrared hygrometer, 120

Kew Observatory net radiometer, 75, 83

Latent heat, 3, 4

Metabolism, 4
Moll-Gorczynski solarimeter, 75, 76–77
Monteith net radiometer, 75, 86

Net radiation
  in forests, 38–41
  in localized energy regimes, 28–41
  in world climates, 21–27
Nusselt number
  defined, 99
  in forced convection, 103, 104, 105
  in free or natural convection, 99–102

Ozone
  absorption bands of, 15, 16, 65, 66
  atmospheric absorption of, 51
  concentration, 14
  radiation from, 18, 66

Photosynthesis, 4
Prandtl number
  defined, 99
  in forced convection, 103, 104
  in free or natural convection, 99, 100

Radiation charts, 67
Radiation fluxes from environments, 69–73
Radiation instruments, 74–93
Radiation laws, 16, 17, 62–64
Reflectivity, 5, 64
  albedo, 60
  of leaves, 138
Relative humidity, 118
Reradiation from a leaf, 72, 131
Reynolds number
  defined, 102
  in forced convection, 103, 104, 105
Robitzsch bimetallic actinograph, 75, 77–78

Scattering, 51
  Rayleigh, 52
  Mie, 52
Schulze radiometer, 75, 90–91
Silver disc pyrheliometer, 75, 86
Soil
  heat wave in, 10, 127, 128
  temperature, 10, 11, 12
Solar radiation, 3, 4, 5, 7, 47–60
  artificial source of, 134–35
  at earth's surface, 5, 28, 29, 49, 50, 52, 54–60
  geographical distribution of, 7, 8
  on slopes, 58, 59
  related to cloud cover, 55, 56, 57
  solar constant, 7, 47
  spectral distribution of, 9, 47–49
Spectral properties
  of atmospheric radiation, 15–18
  of black body radiation, 62
  of leaves, 135–42
  of sunlight, 9, 47–49, 139, 140
Stern pyranometer, 75, 77
Stoll-Hardy radiometer, 75, 86–89
Suomi "economical" net radiometer, 75, 91–92

Temperature
  animal responses to, 41–46
  change of within leaf, 71, 72, 129–33
  in air, 10, 11, 12, 29, 33, 34, 35, 36, 37, 70
  in soil, 10, 11, 12, 128
  of surface, 10, 11, 12, 29, 30, 65, 70
  in trunk or branch, 128, 129
  radiant temperature of sky, 34, 35, 36, 37
  stratospheric, 15, 66
  tropospheric, 15, 66
Temperature profile near a surface, 95
Thermal radiation, 3, 4, 5, 6, 7, 60–69
  detection of, 61
  from atmosphere, 5, 6, 15, 17, 18, 19, 28, 65, 66, 67
  from ground surface, 5, 6, 15, 17, 28, 29, 30
  on animals, 13
  spectral distribution of, 62
Transmission, of atmosphere, 15, 50, 51

Transmissivity, 4
  of leaves, 138
Transpiration, 32, 132–33

Ultraviolet radiation, 51

Vapor pressure, 68, 118
  table of, 69
Viscosity, 96
  of air, 97
  of water, 97

Water vapor
  absorption bands of, 15, 16, 65–66
  atmospheric absorption of, 51
  climatic change due to, 20–21
  concentration, 14, 53, 118
  measurement of, 113–26
  radiation from, 18, 65–66
  total band absorption formulae, 116–17, 121–25
Wave number definition, 115
Wind velocity profile near a surface, 95, 108, 109

3/6

(*continued from front flap*)

still air and in windy air; and the role of transpiration and evaporation in exchanging heat and maintaining the plant or animal at tolerable temperature level.

Instrumentation for solar radiation and for thermal radiation measurements is described in detail and new instrumentation techniques are suggested for water vapor and carbon dioxide observations in ecology. Discussion on other topics includes the rate of heat transfer within a substance, the measurement of leaf temperatures, artificial growth chambers, and the spectral properties of vegetation.

Further depth is added by discussion in terms of the energy-flow concept of various world-wide as well as localized climates.

As complete a bibliography as possible has been given in order that the reader will be encouraged to explore beyond the contents of this book. Further aids to study include thirty-four figures, twelve tables, and a convenient index.

Contents. THE ENERGY ENVIRONMENT IN WHICH WE LIVE. SOLAR AND THERMAL RADIATION. RADIATION INSTRUMENTS. HEAT TRANSFER BY CONVECTION. NEW METHODS IN ECOLOGY. REFERENCES. INDEX.

Printed in Japan

HARPER INTERNATIONAL STUDENT REPRINTS

are new, economy-priced edition of outstanding college and technical texts originally published by Harper & Row. Each is a complete, unabridged, high-quality reproduction of the latest American edition and is generally priced at considerably less than half the cost of the same book in the United States. A complete catalog of Harper International Student Reprints will be mailed on request. Outstanding titles of related interest now published or in preparation include:

Animal Biology	*by Michael F. Guyer & Charles E. Lane*
General Biology	*by Kenoyer, Goddard & Miller*
Introduction to Evolution (2nd Edition)	*by Paul A. Moody*
Nitrogen Metabolism in Plants†	*by George C. Webster*
Principles of Biology	*by W. Gordon Whaley et al.*
Respiratory Metabolism in Plants†	*by Harry Beevers*
Staining Methods: Histologic & Histochemical	*by J. McManus & Robert W. Mowry*
Water Metabolism in Plants†	*by Theodore T. Kozlowski*

†*HARPER REPRINT SERIES IN PLANT PHYSIOLOGY*

For sale in Japan by
JOHN WEATHERHILL, INC.
50 Ryudo-cho, Azabu, Minato-ku, Tokyo

and in other designated countries by
HARPER & ROW, PUBLISHERS
49 East 33rd Street, New York, N. Y. 10016